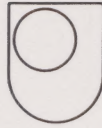




26 Geological Time

27 Earth Materials and Processes





The Open University
Science: A Foundation Course

Unit 27

Earth materials and processes: an introduction to the rock cycle

Prepared by the Science Foundation Course Team

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SCIENCE

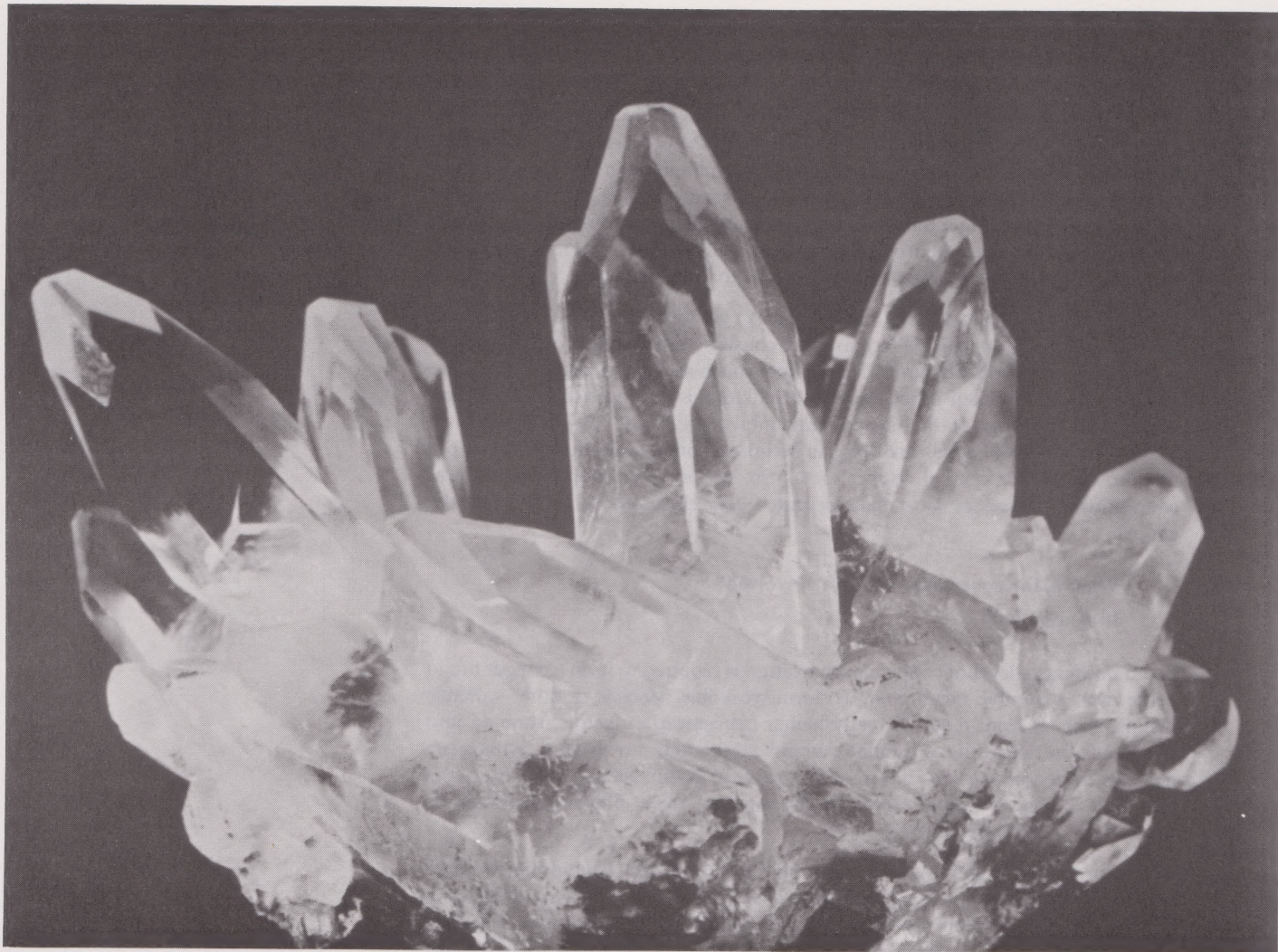
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Table A List of scientific terms and concepts used in Unit 27

| Assumed from general knowledge | Introduced in a previous Unit | Unit No. | Introduced in this Unit | Page No. |
|--------------------------------|--|-------------------|-----------------------------|----------|
| carbon dioxide | anion | 12 | amphibole | 10 |
| conduction of heat | atom | 10 and 11 | andesites | 24 |
| erosion | basalt | 4 (AV tape AC 90) | bed-load | 31 |
| lava | cation | 12 | brittle deformation | 36 |
| viscosity | constructive plate margin | 6 and 7 | calcite | 34 |
| weathering | continental crust | 4 | chemical weathering | 28 |
| | density | 4 | chilled margins | 21 |
| | destructive plate margin | 6 and 7 | clay minerals | 30 |
| | elements | 10 and 11 | cohesion | 33 |
| | gabbro | 4 (AV tape AC 90) | composite cones | 18 |
| | gneiss | SS* | contact metamorphism | 39 |
| | granite | 4 (AV tape AC 90) | creep | 37 |
| | | 6 and 7 | crystallization temperature | 10 |
| | igneous rocks | (AV tape AC 90) | cumulates | 22 |
| | island arcs | 6 and 7 | denudation | 30 |
| | isostasy | 6 and 7 | diapirs | 26 |
| | limestone | 4 (AV tape AC 90) | energy of environment | 31 |
| | mantle | 4 | evaporites | 34 |
| | metamorphic rocks | 6 and 7 | faults | 38 |
| | oxide | 12 | fire fountains | 16 |
| | peridotite | 4 (AV tape AC 90) | folds | 39 |
| | phyllite | SS* | fractional crystallization | 22 |
| | plate tectonics | 6 and 7 | frost shattering | 28 |
| | radioactivity | 10 and 11 | galena | 8 |
| | relationship between stress and strain | 4 | geological cycles | 6 |
| | relative atomic mass | 10 and 11 | heat producing elements | 15 |
| | sandstone | 4 (AV tape AC 90) | hydrated mineral | 10 |
| | schist | SS* | island arc accretion | 25 |
| | | 6 and 7 | joints | 37 |
| | sedimentary rocks | (AV tape AC 90) | laminar flow | 31 |
| | sodium cycle | | mica | 10 |
| | Stokes' law | 26 | mineral | 8 |
| | uniformitarianism | | normal faults | 38 |
| | | | olivine | 9 |
| | | | ooze | 34 |
| | | | partial melting | 20 |
| | | | plagioclase feldspar | 11 |
| | | | plastic deformation | 37 |
| | | | plutonic rocks | 20 |
| | | | physical weathering | 28 |
| | | | pyroclastic rocks | 16 |
| | | | pyroxene | 10 |
| | | | quartz | 10 |
| | | | regional metamorphism | 40 |
| | | | residual minerals | 30 |
| | | | rock cycle | 6 |
| | | | saltation | 31 |
| | | | silicate minerals | 9 |
| | | | sorting | 31 |
| | | | suspended load | 31 |
| | | | tectonic processes | 36 |
| | | | thermal gradient | 40 |
| | | | thrusts | 38 |
| | | | turbulent flow | 31 |
| | | | underplating | |

* Earth Sciences Summer School Notebook.



(Frontispiece) Crystals of calcite. Calcite, CaCO_3 , is an extremely common mineral, forming a large part of common rocks such as limestone. It is also the mineral used by many marine organisms to construct their shells.

Study comment

This Unit is concerned with rocks and how they are made. Reference will be made to the rocks in the Home Kit that you studied first when reading Unit 4. It would be a good idea to re-familiarize yourself with these rocks, perhaps by working through Band II of tape AC 90. A new group of rocks, the metamorphic rocks, will be discussed in this Unit, so you should certainly work through Band III of AC 93 before reading Section 5.

In this Unit, some of the theoretical implications of the Home Experiment that you carried out as part of Unit 26 are discussed, so you should ensure that you have completed this experiment before reading Section 4.

1 Introduction

Unit 26 covered some important concepts connected with geological time and its measurement. In this Unit, you will study some of the processes that form rocks. As you travel around the country, you may get the impression that the rocks you see are rather static, unchanging things. That is because anything to do with rocks generally takes place on a timescale which is quite different from our own. Rocks are in fact being formed all around us, now, and all the time. If we lived for 70 000 years, rather than 70, it would be easier for us to appreciate this.

You probably have a handful of small change in your purse or pocket. Consider a ten-pence piece. It weighs 11 grams, and consists of a mixture of copper and nickel. Now you probably don't value copper much as a metal, and wouldn't lose any sleep if you lost a ten-pence piece. But think of this: the average concentration of copper in the rocks of the Earth's continental crust is only 45 *parts per million*. The lowest grade of copper ore that can be mined contains about 4000 parts per million, or 0.4 per cent of copper metal. So, to form the copper ore, natural concentration processes must have operated very efficiently to concentrate the copper.

What is the concentration factor involved?

It is about 100.

So an impressive level of copper concentration is required to form a copper ore to make even a humble metal coin. To form workable deposits of other valuable metals, such as lead, requires vastly higher concentration factors, running into thousands. In this Unit, we cannot examine all the steps involved in forming ore deposits from crustal rocks, but we shall be able to examine many of the processes that lead in that direction, especially the ways in which different rocks can be made from a common starting point.

The starting point is peridotite. As you will recall from Unit 4, peridotite forms 80 per cent by volume of the Earth. There is, therefore, clearly a lot of it and, as you will see later in this Unit, it is thought that almost all crustal rocks are derived from mantle peridotite. Your sandstone specimen S6 doesn't look anything like peridotite and, indeed, it is chemically totally dissimilar, but it was derived originally from peridotite. To derive the one rock from the other is by no means a simple, single stage process. To achieve it requires aeons of time, and endless repetitions of the same sorts of process. In fact, to make a sandstone or an orebody requires that rocks go through *geological cycles*.

1.1 Geological cycles

Geological cycles act within the Earth on large and small scales, over long periods and short ones. They may affect large masses of rock, or single elements; they may be simple or complex.

Can you think of a major geological cycle acting in a plate tectonic context?

You should recall that when basaltic oceanic crust is created at a constructive plate margin, it returns to the mantle again at a destructive plate margin. This is an example of a geological cycle acting on a very large scale; it is extremely important in the manufacture of rocks, as we shall see.

Another important geological cycle is the sodium cycle, which you were introduced to in Unit 26; another is the *water cycle*, or *hydrological cycle*, in which rain-water falling on land is carried to the sea in rivers, evaporates, condenses, and falls again over land areas as rain, hail or snow.

Intimately linked to the hydrological cycle is the best-known cycle in geology, sometimes called the *rock cycle*. This was first recognized by the pioneer Scottish geologist James Hutton. In a book published in 1785 called, rather immodestly, *Theory of the Earth with Proof and Illustrations*, he showed how igneous rocks may be eroded to form sediments by weathering and decay; the sediments become compacted into rocks, and a later mountain-building event then exposes these

hydrological cycle

rock cycle

sedimentary rocks at the surface, where they are eroded away to form a fresh generation of sediments (Figure 1).

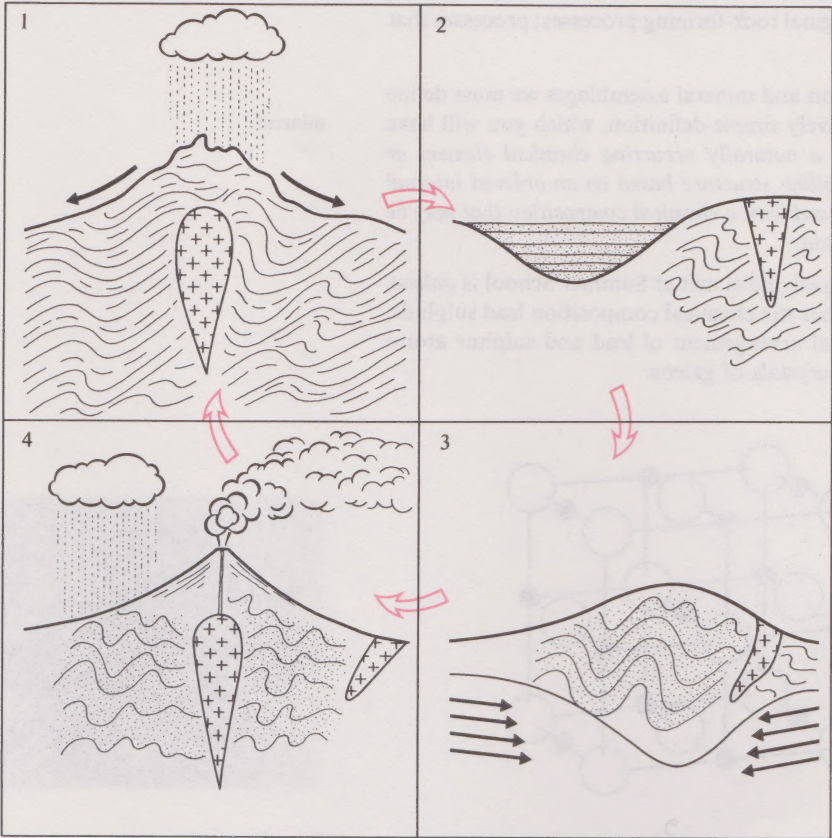


FIGURE 1 The rock cycle, as envisaged by Hutton: (1) Existing rocks are eroded; (2) deposited to form sediments; (3) thrown up by earth movements into a mountain belt; (4) subjected to igneous and metamorphic processes; and eroded once more.

A cycle of any kind requires some kind of energy source to drive it. What are the energy sources driving the hydrological and rock cycles?

The hydrological cycle is relatively easily explained, because the energy that drives it comes from the heat of the Sun. The rock cycle is more complex. Part of it is wrapped up in the hydrological cycle, because moving water plays a major part in the formation of sedimentary rocks. Equally important, however, is the Earth's *internal* heat, which is responsible for driving plate tectonic processes and for building up the mountain belts, which are subsequently eroded. We shall be considering these different components of the rock cycle in later Sections of this Unit. First, though, we must look more closely at the minerals of which rocks are made.

2 Rock materials

2.1 Rocks and minerals

Have you ever stopped to consider precisely what is meant by a 'rock'? This word has been part of your vocabulary since childhood, but it is unlikely that you have ever been asked to define it. You have already examined some of the rocks from your Home Kit, and others at Summer School. It was demonstrated in the Audio-visual tape for Unit 4 that rocks are composed of individual grains that are either cemented together, or crystallized into an interlocking mass. These grains are *mineral* grains, and almost all rocks are simply consolidated aggregates of mineral grains.

Igneous rocks, which form by crystallization from a silicate melt, actually contain several different kinds of minerals, as you can see for yourself in Home Kit specimens S1 and S5. Some sedimentary rocks, however, may be composed predominantly of a single mineral, reflecting some process of selection in the rock

cycle that has favoured its inclusion. Specimen S6, for example, is made up almost entirely of *quartz* grains, and specimen S7 contains a high proportion of the mineral *calcite*. The assemblage of minerals found in a rock is not, therefore, random but directly related to the original rock-forming processes; processes that we shall be exploring in this Unit.

Before we can study mineral formation and mineral assemblages we must define what we mean by a mineral. A relatively simple definition, which you will have already met at Summer School, is: *a naturally occurring chemical element or compound, possessing a definite crystalline structure based on an ordered internal arrangement of the constituent atoms, and with a chemical composition that may be expressed in terms of a chemical formula.*

An example of a mineral you may already have met at Summer School is *galena*. Galena is a silvery grey mineral that has the chemical composition lead sulphide, PbS. Figure 2 shows how the internal arrangement of lead and sulphur atoms gives rise to the typical cube-shaped crystals of galena.

mineral

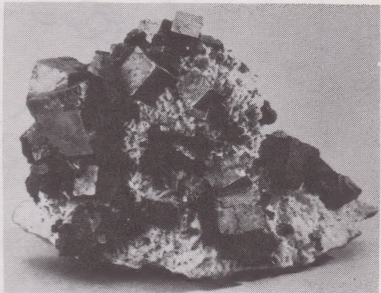
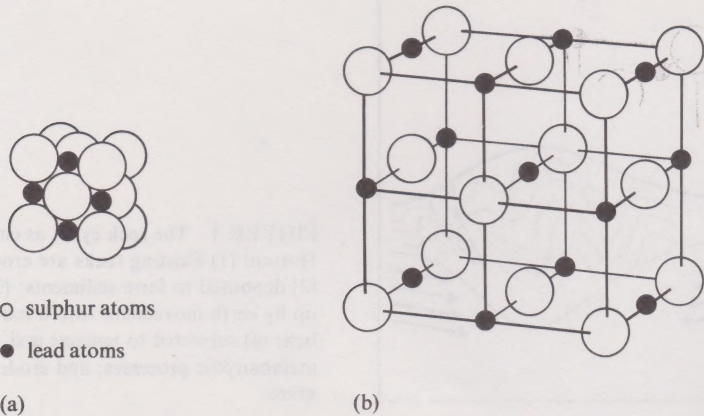


FIGURE 2 (a) The packing of the atoms of lead and sulphur in the mineral galena. The relative sizes of the atoms are shown. (b) The atomic structure of galena to show the way in which the atoms are linked to produce a cubic structure. (c) Cubic crystals of galena.

ITQ 1 Given below are some common substances, and their chemical compositions. Applying the definition of a mineral given above, which would you classify as minerals? (You may assume that they all have definite crystalline structures.)

- | | |
|-------------------------|-------------------------|
| (a) Ruby | Al_2O_3 |
| (b) Cooking salt | NaCl |
| (c) Industrial diamonds | C |
| (d) Snowflakes | H_2O |

More than 2 000 mineral species and varieties of mineral species are now known, and new ones are discovered annually. These are classified into groups based on similarities in chemical composition and atomic structure. If you have a hand-sized specimen of a coarse-grained rock, it is possible to distinguish between the major mineral groups, and sometimes between individual minerals, because each group possesses measurable physical and chemical properties that are directly related to the atomic structure and chemical composition of the group. Some of these properties you will have already investigated, such as colour, cleavage, hardness and reactivity with acids, and they are also explained fully in your Earth Science Summer School laboratory notebook. The recognition of minerals in fine-grained rocks, however, and the identification of rare minerals, are much less simple and may require the use of a microscope or X-ray techniques, which we cannot explain here.

Table 1 lists the abundances of some important elements in the Earth's continental crust.

Which two elements in this Table are by far the most abundant?

Oxygen and silicon are much more abundant than any other elements. Silicon is three times as abundant as aluminium, the third most common element overall.

Since oxygen and silicon are so overwhelmingly abundant, it will come as no

surprise to you to learn that all the most common rock-forming minerals are *silicates*. Fortunately, although they occur in huge amounts, there is only a handful of different kinds of silicate minerals. We examine these in the next Section.

TABLE 1 The composition of the Earth's crust

| (a) | | | | (b) | | |
|--|----|----------|-------------------|---|----|-------------------|
| Abundances of the most common elements by weight | | | | Abundances by weight of some elements used in everyday life | | |
| | | per cent | parts per million | | | parts per million |
| oxygen | O | 46.6 | 466 000 | sulphur | S | 520 |
| silicon | Si | 27.7 | 277 000 | carbon | C | 320 |
| aluminium | Al | 8.3 | 83 000 | chlorine | Cl | 200 |
| iron | Fe | 5.0 | 50 000 | nickel | Ni | 80 |
| magnesium | Mg | 2.1 | 21 000 | zinc | Zn | 65 |
| calcium | Ca | 3.6 | 36 000 | nitrogen | N | 46 |
| sodium | Na | 2.8 | 28 000 | copper | Cu | 45 |
| potassium | K | 2.6 | 26 000 | lead | Pb | 15 |
| TOTAL | | 98.9 | | tin | Sn | 3 |
| | | | | uranium | U | 2 |
| | | | | mercury | Hg | 0.5 |
| | | | | silver | Ag | 0.1 |
| | | | | gold | Au | 0.005 |

2.2 The silicate minerals

The atomic structure of this mineral group is based upon the silicate unit $(\text{SiO}_4)^{4-}$ (Figure 3), which can be represented as a sort of tetrahedral building block. (A tetrahedron is a four-sided body, like a pyramid.) The unit has four residual negative charges that are balanced either by appropriate cations such as Fe^{2+} , Mg^{2+} , Na^+ , K^+ and Al^{3+} , or by the sharing of one or more oxygen atoms with other silicate units. The different ways in which the tetrahedral silicate building blocks are linked together, and the choice of cations to balance the negative charges, are responsible for the variety of silicate mineral species.

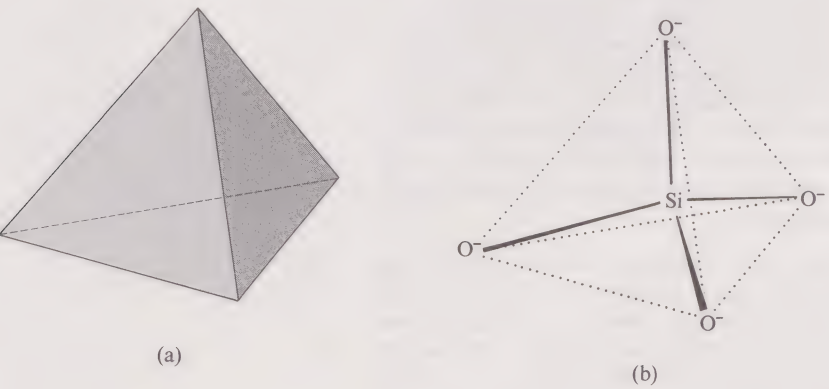


FIGURE 3 (a) The tetrahedron—the basic shape of the $(\text{SiO}_4)^{4-}$ unit. (b) The atomic structure of the $(\text{SiO}_4)^{4-}$ unit to show how the silicon and oxygen atoms are linked.

Six major sub-groups of silicate minerals are recognized, based upon the way in which the silicate units are joined to each other by sharing of oxygen atoms, and these, together with their chemical compositions, are listed in Table 2. The elements that are grouped together in round brackets are able to occur in place of each other. Thus, for example, the olivines may contain *all* magnesium (Mg_2SiO_4) or *all* iron (Fe_2SiO_4), or a mixture of the two in any proportion. Some of these minerals should be familiar to you, as you will have seen them in the Earth Sciences laboratories at Summer School.

TABLE 2 The structure and chemical compositions of the most common sub-groups of silicate minerals in rocks

| | Sub-group | Most common colour in rocks | Chemical composition | Structure— oxygens shared per silicate unit | Temperature of crystallization |
|--|------------|--|---|---|-----------------------------------|
| Iron- and magnesium- bearing minerals | OLIVINES | pale green (Mg-rich) to yellowy-brown (Fe-rich) | $(\text{Mg, Fe})_2\text{SiO}_4$ (the Mg-rich variety is the gemstone peridot) | olivine 0 | ↑ HIGHEST |
| | PYROXENES | dark green to black | $(\text{Mg, Fe, Ca})_2\text{Si}_2\text{O}_6$ | pyroxene 2 | |
| | AMPHIBOLES | dark green to black | a hydrated aluminium-silicate containing Ca, Na, Mg and Fe | amphibole $2\frac{1}{2}$ | |
| | MICAS | brown, (Fe-rich) biotite to white, (Fe-poor) muscovite | a hydrated aluminium-silicate containing K, Mg, Fe a hydrated aluminum-silicate containing K | micas 3 | |
| Iron- and magnesium- poor minerals | FELDSPARS | white or sometimes pink—plagioclase | $\text{CaAl}_2\text{Si}_2\text{O}_8$ and $\text{NaAlSi}_3\text{O}_8$ (or any mixture of Ca and Na) | feldspars 4 | ↓ LOWEST |
| | | white or pink— potassium-feldspar | KAlSi_3O_8 | | |
| | QUARTZ | Colourless | SiO_2 | quartz 4 | |

Table 2 shows that no oxygens are shared between silicate units in the olivines; as one proceeds down the Table, more and more oxygen sharing takes place. This produces a more open atomic structure and so large cations such as K^+ , Ca^{2+} and Na^+ are found in amphiboles, micas and feldspars. In some cases, such as the amphiboles and micas, there are some gaps in the structure that are just the right size for hydroxyl ions, OH^- , to be accommodated. These minerals are, therefore, said to be *hydrated*.

hydrated minerals

Do you notice a difference between the colours of the minerals rich in iron and magnesium, and those poor in these elements?

The iron- and magnesium-rich minerals are all green, brown or black, whereas the others are all pale coloured, white or colourless, because it is the presence of iron that makes minerals dark.

2.3 The crystallization of the silicate minerals

Table 3 lists the rough chemical compositions of some common crustal rocks, similar to those in your Home Kit. Each rock is made up of one or more different kinds of silicate minerals. Don't be overawed by all the numbers in this Table, you will not have to recall them. Notice, too, that instead of simply listing the percentages by weight of each element, the Table is expressed in terms of element *oxides*. This is conventional in the Earth sciences for historical reasons. It is most important that you should *not* conclude that elements are actually present as oxides in all cases. Some oxides such as TiO_2 are indeed present in that form, but others, such as K_2O would be violently unstable chemically, and the potassium is always bound up in more stable compounds.

You now have some chemical data on rocks, and on the minerals that make up rocks. By inspecting these two sets of data, you might get the impression that you could arrive at *any* of the rock compositions given, simply by mixing different minerals in the right proportions. Thus, by juggling numbers, you might be able to arrive at a granite composition by mixing quartz, olivine, pyroxene and feldspars. The fact is, however, that this simple mixing does *not* take place, and different rocks are always characterized by distinctive minerals. Peridotites, for example, invariably contain olivine and pyroxene, whereas olivine is *never* found in granites.

The physical factor that controls the presence of different minerals in rocks that cool from molten material, or *magma*, is the *crystallization temperature*. Each mineral is characterized by a different crystallization temperature, hence, when a magma begins to cool, the silicate minerals will begin to crystallize in a sequence that is controlled by these temperatures. The approximate order in which silicate

crystallization temperature

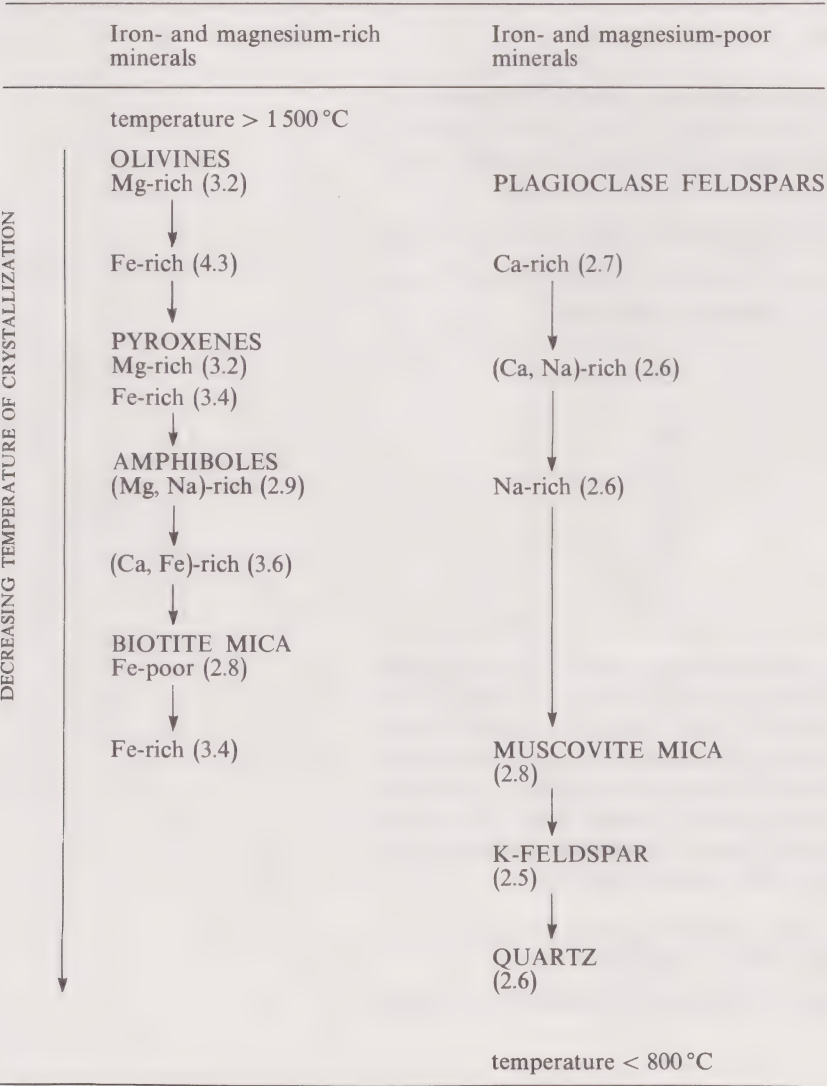
TABLE 3 The chemical composition by weight (%) of some common crustal rocks (comparable to those in the Home Experiment Kit)

| | Peridotite | Basalt | Andesite | Granite | Sandstone |
|---|------------|--------|----------|---------|-----------|
| SiO ₂ | 45.7 | 49.2 | 56.0 | 71.8 | 81.0 |
| TiO ₂ | 0.13 | 1.4 | 0.7 | 0.4 | — |
| Al ₂ O ₃ | 2.7 | 15.8 | 16.4 | 14.6 | 7.6 |
| Fe ₂ O ₃ + FeO | 7.3 | 9.4 | 7.8 | 3.4 | 4.3 |
| MgO | 41.5 | 8.5 | 6.4 | 0.9 | — |
| CaO | 2.0 | 11.1 | 7.9 | 2.0 | — |
| Na ₂ O | 0.22 | 2.7 | 2.9 | 3.5 | 0.6 |
| K ₂ O | 0.03 | 0.26 | 1.6 | 4.2 | 4.8 |
| TOTALS* | 99.6 | 98.4 | 99.7 | 100.8 | 98.3 |

* The totals of the analyses are not 100 per cent because several minor constituents have been omitted from this Table.

minerals crystallize is shown in Table 4. This tabular summary represents a considerable simplification of the complex natural situation. The order of crystallization can be affected by other physical factors, such as pressure, and by chemical factors. Furthermore, the sequence is a generalized one; it is not intended to summarize the minerals that will crystallize when any particular rock melt is cooled.

TABLE 4 The sequences of crystallization of the common silicate minerals as a function of temperature. (The figures in brackets are the densities in 10³ kg m⁻³.)



We have indicated the order in which each mineral group begins to crystallize in both Tables 2 and 4. You can see that the Ca-bearing plagioclase feldspars are able to crystallize at the same time as some of the iron- and magnesium-rich minerals. However, mica, K-feldspar and quartz are generally among the last minerals to crystallize.

From the data given in Tables 2 and 4 and the questions you have answered, you should now be able to see how many of the properties of the silicate minerals are related to each other, and how they vary with decreasing temperature of crystallization. For example, you have already seen that the *colour* of a mineral is affected by its *chemical composition*, and that chemical composition is *controlled by the structure* of a mineral. You should now be able to work out how *chemical compositions and structure* are also related to the *temperature of crystallization* of a mineral.

- ITQ 2** (a) How do the proportions of magnesium and iron change with temperature *within* any individual mineral group such as the olivines?
- (b) How do the proportions of calcium and sodium change with temperature *within* the plagioclase feldspar group?

Structurally, the high-temperature iron- and magnesium-rich minerals tend to form single $(\text{SiO}_4)^{4-}$ units, with *no* shared oxygens. At lower temperatures, progressively more oxygen-sharing takes place until *all* the oxygens in an $(\text{SiO}_4)^{4-}$ are shared with adjacent units.

The control of the order of their crystallization by temperature, and the structures of the silicate minerals, both have important consequences when we consider how igneous rocks break down to form the raw materials for sedimentary rocks (Section 4.0).

2.4 The densities of silicate minerals

We can now begin to make some deductions about the *densities* of minerals, and consequently the densities of different types of rocks. Table 5 lists in decreasing order the approximate relative atomic mass of the elements that make up the silicate minerals.

TABLE 5 The approximate value of the atomic weights of the common elements of silicate minerals

| Element | Relative atomic mass |
|----------------|----------------------|
| iron (Fe) | 56 |
| calcium (Ca) | 40 |
| potassium (K) | 39 |
| silicon (Si) | 28 |
| aluminium (Al) | 27 |
| magnesium (Mg) | 24 |
| sodium (Na) | 23 |
| oxygen (O) | 16 |

When you compare the atomic mass of iron and magnesium, it is quite easy to see why the iron-rich members of any mineral group are denser than the magnesium-rich members of the same group. Similarly, if you compare the relative atomic mass of calcium and sodium, it is clear why calcium-rich plagioclase feldspars are more dense than sodium-rich varieties. It is much more difficult to interpret differences in densities between *different* mineral groups. Why, for example, should magnesium-rich olivine be more dense than calcium-rich plagioclase, despite the comparatively high relative atomic mass of calcium?

- What factor other than the relative atomic mass of its constituent atoms should affect the density of any crystalline material?
- Clearly, the more closely packed the atoms are in a structure, the denser that material will be.

You have already seen that there is more oxygen sharing in feldspar structures than in olivines, and this leads to a more open structure. Thus it is the *closer packing of atoms* in the magnesium-rich olivine structure that makes it more dense than calcium-rich plagioclase. Rocks that are made up of high-density minerals naturally have high densities themselves.

2.5 The distribution of silicate minerals in crustal rocks

So far, we have considered the major groups of silicate minerals that occur in igneous rocks. We must now turn to the two other great groups of rocks, the sedimentary and metamorphic rocks. You will already know something about the minerals in sedimentary rocks from the Audio-visual sequence for Unit 4 but, so far, little has been said about metamorphic rocks. They are simply rocks that have experienced changes; they may be either igneous or sedimentary initially, but have been subjected to heat and pressure to such an extent that they are recognizeably different from their parent rocks. In extreme cases, it is often impossible to determine the origin of a metamorphic rock.

Table 6 lists the main groups of silicate minerals found in igneous, sedimentary and metamorphic rocks.

TABLE 6 The silicate mineral groups found in igneous, sedimentary and metamorphic rocks

| Igneous rocks | Sedimentary rocks | Metamorphic rocks |
|---------------|-------------------|-------------------|
| olivines | clay minerals | garnets |
| pyroxenes | | |
| amphiboles | | amphiboles |
| micas | micas | micas |
| feldspars | K-feldspars | feldspars |
| quartz | quartz | quartz |

Two of these mineral groups, clay minerals and garnets, have not been discussed before. Clay minerals, as their name suggests, are the chief constituent of clays. They are hydrated aluminium silicates, somewhat similar to micas in their structure. Garnets are familiar as gemstones and as industrial abrasives. They are a large group, structurally somewhat similar to the olivines.

Look carefully at Table 6, and then answer these questions:

What are the principal differences between the minerals of the three groups? Are the minerals of the igneous rocks more like those of the metamorphic rocks than those of the sedimentary rocks? If so, why?

The olivines are found only in igneous rocks, whereas clay minerals are found only in sediments, and garnets only in metamorphic rocks. Amphiboles occur both in igneous and metamorphic rocks, but not in sediments. There are clearly more points of similarity between the metamorphic and igneous rocks than between sedimentary and igneous rocks. The reasons for the differences in mineral assemblages between the major rock types are fundamental, and are due to the operation of different processes within the rock cycle. The minerals of igneous rocks are formed at high temperatures. On erosion, they are exposed to quite different physical conditions. Some minerals are mechanically broken down into fragments, and others are chemically altered. The products of both processes are involved in the formation of sedimentary rocks, which takes place at the surface of the Earth, at low temperatures. When sediments are metamorphosed, they are heated up, pre-existing minerals may be destroyed, and new ones formed. Since the physical conditions of formation are much more like those of igneous rocks than sedimentary ones, it is only to be expected that metamorphic minerals show similarities to igneous ones. We shall be studying the important component parts in the rock cycle that lead to the formation of the major rock groups in Sections 3, 4 and 5.

Objectives for Section 2

Now that you have completed Section 2 of this Unit you should be able to:

- (a) Indicate that you have some idea of the meaning of the terms: mineral, silicate minerals, crystallization temperature, atomic structure. (Some of this is tested in ITQ 1 and SAQ 1)
- (b) Relate the colour of a silicate mineral to its chemical composition, and the chemical composition to structure and to the temperature of formation of the mineral. (ITQ 2 and SAQ 1)
- (c) Relate the chemical composition and structure of silicate minerals to their densities. (SAQs 1 and 2)
- (d) Make predictions about the mineral composition, temperature of formation, order of mineral crystallization and density of an igneous rock from a hand specimen. (SAQs 1 and 2)

SAQ 1 Select from your Home Experiment Kit the coarse-grained samples S1 (granite), S4 (peridotite) and S5 (gabbro), and make sure that you have a strong light to shine on the samples and a hand-lens or magnifying glass available to see the individual minerals more clearly.

- (a) Table 7 lists the three igneous rocks you have selected, and the average mineral compositions for these rocks. Using your hand specimens, fill in the empty spaces in the Table 7, and then try and answer the questions below. (Remember that the specimens will vary from sample to sample, and that the mineral compositions quoted are averages for that rock type. You should complete the Table—from your own observations, and note whether they differ slightly from the completed Table in the answer to this SAQ. You should not worry about minor discrepancies.)
- (b) Why is there an overall colour difference between rock samples S1 and rock samples S4 and S5?
- (c) What are the reasons for the density differences between samples S1, and S4 and S5?
- (d) The feldspar crystals in sample S1 have good, rectangular crystal outlines, whereas the quartz crystals are irregular. Can you explain why this might be so?
- (e) What can you say about the *rate* of cooling and crystallization of all three samples? (Remember the comments from the Audio-visual sequence!).

TABLE 7 (for use with SAQ 1 and SAQ 2)

| Rock sample | Average mineral composition | Colour of mineral | Overall rock colour (light or dark) | Relative density (high, moderate low) | Relative temperature of formation (high, moderate, low) | Anticipated sequence of mineral crystallization |
|---------------|--------------------------------|-------------------|-------------------------------------|---------------------------------------|---|---|
| S1 GRANITE | 10% biotite mica | | | | | |
| | 60%, mainly Na- and K-feldspar | | | | | |
| | 30% quartz | | | | | |
| S5 GABBRO | 10% olivine | | | | | |
| | 30% pyroxene | | | | | |
| | 60% Ca- and Ca-Na-feldspar | | | | | |
| S4 PERIDOTITE | 85% olivine | | | | | |
| | 15% pyroxene | | | | | |

- SAQ 2** Take the basalt sample S3 from your Home Experiment Kit:
- (a) From the overall *colour* of the rock, can you suggest which types of minerals you would expect to be present?
 - (b) Can you suggest what the *relative* temperature of formation of S3 was: high, moderate or low?
 - (c) What can you say about its rate of cooling and crystallization?
 - (d) You have measured the densities of samples S1 (granite), S3 (basalt) and S4 (peridotite). Which of the samples in Table 7 is S3 closest to?

Note It is important that you read the answers to SAQ 2 before continuing to the next Section.

3 Igneous processes

Igneous rocks form the starting point in the rock cycle: the materials making up all sedimentary and metamorphic rocks were derived originally from an igneous source, although for any given rock this primitive source may be lost in the mists of time. In this Section, we shall be examining the essential part that igneous rocks play in the rock cycle, but we shall also investigate how igneous rocks themselves are formed, and try to answer some important questions such as:

- Why is there such a wide range of igneous rocks?
- How can one make a granite from a basalt?
- Why are quite different kinds of lava erupted from volcanoes at constructive plate margins from those erupted at destructive plate margins?

The answer to almost all problems in geology rest in the mantle, so it is there that we shall start.

3.1 The mantle

Eighty per cent of the Earth’s volume consists of mantle rocks. Arguably, therefore, we ought to devote most of this Unit to it! Unfortunately, however, the mantle is inaccessible, so it is impossible to say much about it—we have to rely on inferences drawn from surface observations of rocks and minerals.

You will recall from Unit 4 that peridotite is the main rock of which the mantle is made, similar to your hand specimen S4. The chemical composition of the peridotite samples found at the surface of the Earth has been exhaustively studied; as you have already learned, peridotite is composed mainly of the mineral olivine, an iron- and magnesium-rich silicate. Now you will recall from Units 6 and 7 that the process of sea-floor spreading is driven by convection currents in the mantle, which rise up and spread outwards beneath the mid-ocean ridges.

But what is the source of heat that ‘drives’ these convection cells?

By far the most likely source of the Earth’s internal heat is decay of long-lived radioactive isotopes of the elements K, U and Th. Since the mantle is thought to be in a state of constant convective motion, you might suppose that it is rich in these radiogenic elements. Not so. In fact, peridotites contain only minute traces of the radioactive isotopes.

heat-producing elements

TABLE 8

| Rock | Abundance of radioactive element/ppm | | | Heat produced/ J kg ⁻¹ s ⁻¹ |
|------------|--------------------------------------|---------|-----------|--|
| | uranium | thorium | potassium | |
| granite | 4 | 20 | 40 000 | 9.5 × 10 ⁻¹⁶ |
| basalt | 0.5 | 2 | 15 000 | 1.8 × 10 ⁻¹⁶ |
| peridotite | 0.02 | 0.06 | 200 | 3.2 × 10 ⁻¹⁸ |

Table 8 shows just what poor producers of heat peridotites are especially when compared with granite. This fact has encouraged speculation that the Earth's internal heat and the energy to drive mantle convection currents, have a different source.

One possibility is that the Earth still retains a part of its *original* heat, trapped inside it when it was formed. Rocks are such extremely poor conductors of heat that heat contained in the core at the time of the Earth's formation, over 4 billion years ago, would not all have leaked away into space. Another possibility is that the Earth's core may contain a significant amount of potassium, and that decay of the radioactive isotope of potassium in the core would contribute the heat necessary to generate convection cells at the core-mantle boundary.

As you saw in Unit 4, however, it is generally supposed that the non-metallic element in the core is sulphur, which does not have radioactive isotopes. It seems that whereas the peridotite of the mantle is relatively poor in heat-producing elements, there is such a large mass of it that enough heat can be generated within it to keep it at a high temperature.

3.2 Volcanoes and volcanic rocks

One of the most obvious consequences of the Earth's hot, mobile interior is volcanic activity at the surface. You will already have seen something of volcanic rocks in the Audio-visual sequence of Unit 4 and TV programmes 5-7. Volcanic rocks scarcely need any definition—they are the products of volcanoes. But they have three important characteristics which you should recall:

- 1 They crystallize at the surface of the Earth.
- 2 They are usually fine grained.
- 3 They do *not* cut across older rocks, but rest on top of them.

Volcanic rocks come in a wide range of shapes and forms. Your Home Kit specimens S2 and S3 are both volcanic rocks, basalts, but one, (S3) was erupted as liquid lava and the other (S2) as a bubbly, low-density fragment that was hurled into the air in a volcanic eruption and subsequently fell to Earth to form part of a cindery, ash-like deposit. Such fragmented igneous rocks are known as *pyroclastic* (fire-broken) rocks. Apart from their marked *physical* differences, however, S2 and S3 are almost identical in *chemical* composition. Volcanic rocks of many other different compositions are known, and these may occur either as lavas or as pyroclastic rocks.

pyroclastic rocks

When a basalt magma is first erupted, it is at a temperature of about 1 100 °C. At this temperature, its viscosity is fairly low, so that it can flow and slosh around in a very liquid fashion. As it cools, it becomes stiffer and more viscous. In this respect, basalt magmas are similar to almost all other liquids, which get progressively more viscous as they cool. You will certainly have noticed how sluggish your car engine is on a crisp, winter morning, when the engine oil is thick and viscid.

When a high temperature, low-viscosity basalt magma approaches the surface, the dissolved gases pent up within it come out of solution like the bubbles in a beer bottle when the top is taken off. Because the magma is of low viscosity, the gas can blow off freely. The escaping gas often sprays showers of liquid basalt into the air as *fire fountains*; thus, generally speaking, basaltic eruptions are not violently explosive, and there is more lava erupted than ash.

The low viscosity of basaltic magma controls the shape of the volcanic structures that it builds up. Basalt lavas spread out widely around the vent from which they are erupted—this property is beautifully displayed on the Moon, where basalt lavas were also produced by melting in the lunar mantle (Figure 4).

Because they spread out so widely, basaltic lavas accumulate to form broad, gently sloping volcanoes. The best-known examples of these are in the Hawaiian islands. Although the slopes on these islands rarely exceed a few degrees, they are the biggest volcanoes in the world and, taking account of the part under water, they would also be the highest *mountains* in the world, since they rise over 9 km from the sea-floor. (Everest is only 8.8 km high.)

This raises an interesting question. How is it that a volcano can build itself up so

high at all? The answer lies in the density of rocks once more. Let us consider a basaltic volcano erupting on the floor of the ocean, where the depth to the low-velocity layer is 50 km. The low-velocity layer, you will recall, is the partially molten layer within the mantle that allows plate tectonic movements to take place.

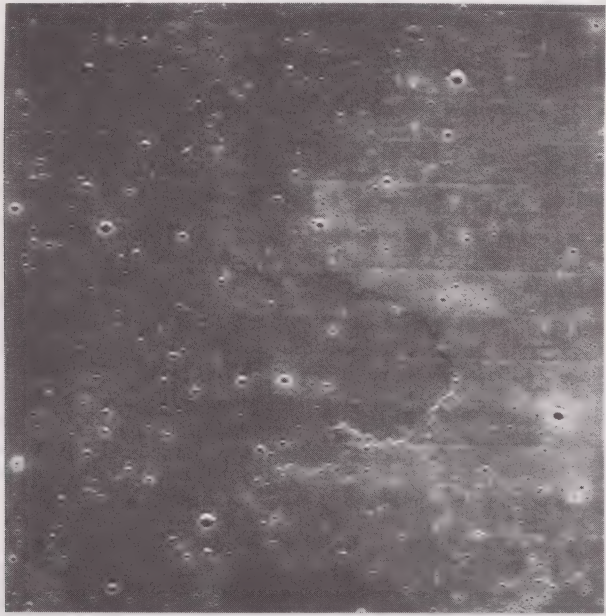


FIGURE 4 Basalt lavas on the surface of the Moon. Basalts form all the grey areas or 'seas' which make up the 'Man in the Moon'. The front of the basalt lava flows can be traced around the irregular, lobate, dark-grey area near the centre of the picture. (NASA Lunar Orbiter V)

Assuming that the density of the lithospheric plate is $3.3 \times 10^3 \text{ kg m}^{-3}$, what would the mass of a vertical column of one square metre cross-sectional area be?

It would be $1 \times 50 \times 10^3 \times 3.3 \times 10^3 \text{ kg}$ (area times height times density) = $165 \times 10^6 \text{ kg}$.

Now a typical basalt has a density of about $2.9 \times 10^3 \text{ kg m}^{-3}$, as you have found from the Home Experiment completed in Unit 4.

What height, therefore, would a metre-square column of basalt have to be to have the same mass as the lithospheric column? (You should make the simplifying assumption that the liquid basalt has the same density as the solid, which is not strictly true in Nature.)

Simply divide the mass by the density:
$$\text{height} = \frac{165 \times 10^6}{2.9 \times 10^3}$$
$$= 57 \text{ km}$$

The basalt column would be 7 km higher than the lithospheric one. In practical terms, this means that a basaltic volcano can keep on building itself upwards until the pressure of the magma column in its vent exactly balances the pressure due to the lithospheric plate around it (Figure 5). Once it reaches this magic figure, it can no longer grow vertically.

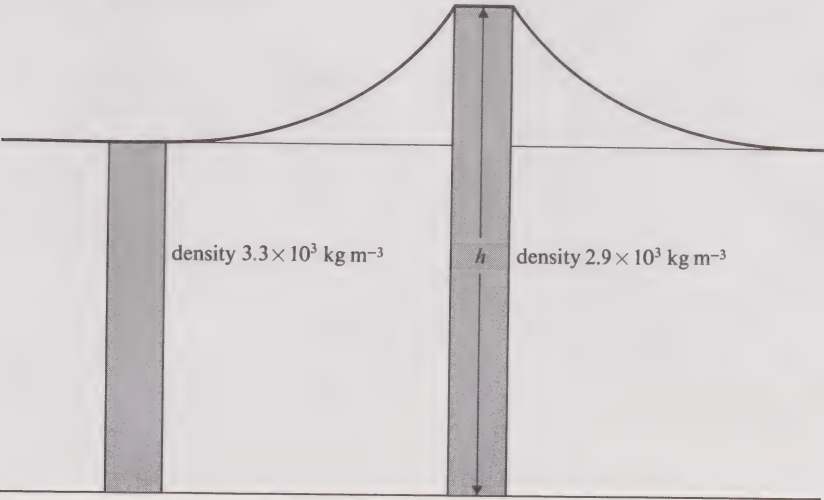


FIGURE 5 The height of a volcano h is obtained by balancing the pressure due to its magma column against that of the pressure of a similar column of the denser surrounding material.

But what will happen next? Clearly, if magma is still available, it will break out at a lower level and build up a new volcanic structure. Most of the volcanic islands in the world display this situation—there is usually one active volcanic centre, and one or more extinct ones.

Can you think of another factor that may affect the height of a volcano?

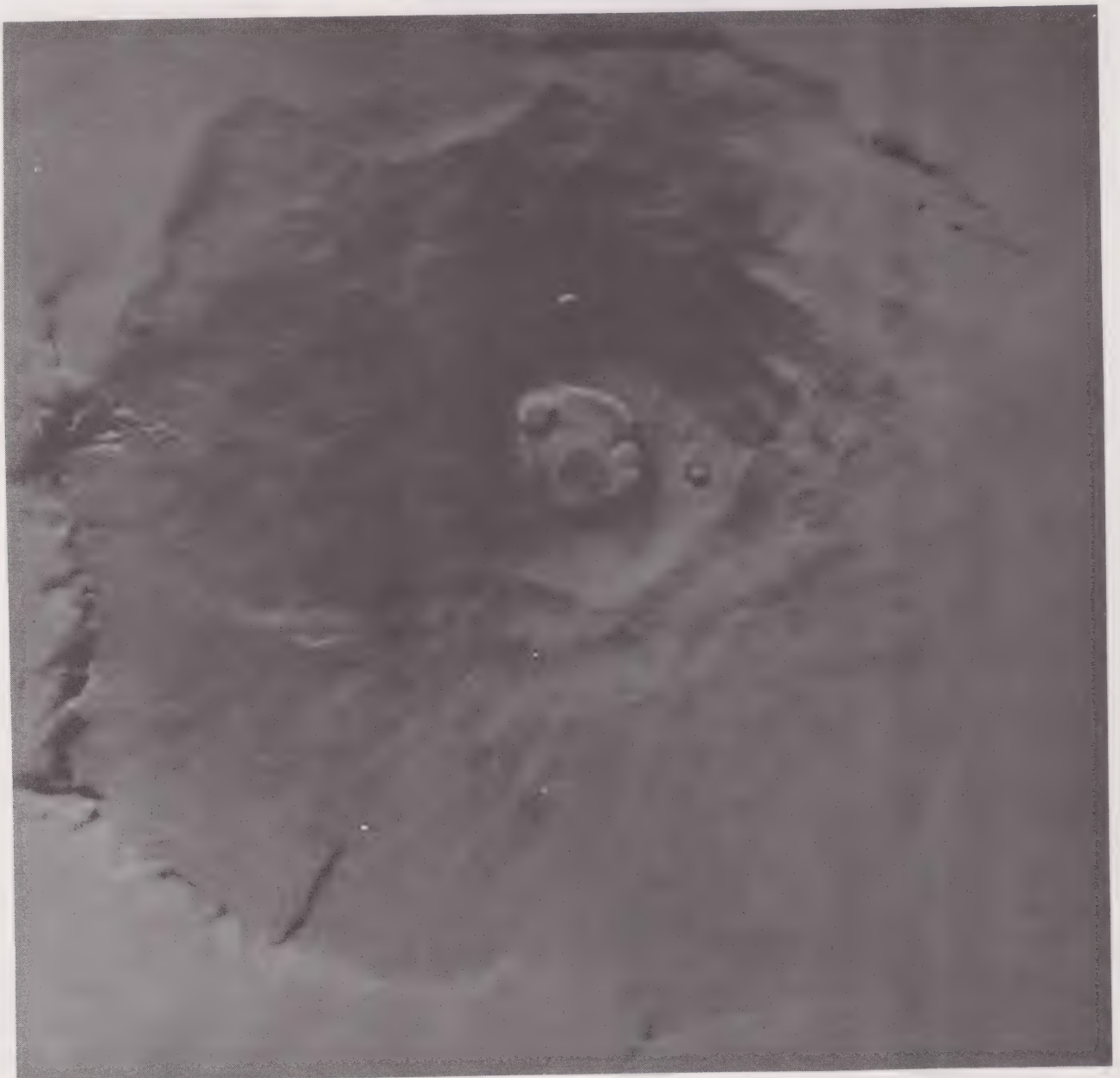
You should recall the effect of isostasy (Units 6 and 7, Section 4.4). A 3000-m-high volcano may be constructed in only a million years or so, and clearly represents a large local excess of mass. Because it is built up more rapidly than isostatic readjustment can compensate for, a positive gravity anomaly will exist. With the passage of time, the volcano will sink slightly down into the mantle and the gravity anomaly will disappear, as you saw in Units 6 and 7, Figure 21.

ITQ 3 How high would a volcano grow above a lithospheric plate 60 km thick, with a density of $3.2 \times 10^3 \text{ kg m}^{-3}$? (You may assume that the basalt magmas have a density of 2.9.)

Although the Hawaiian volcanoes are impressively high, they are goose pimples compared with those on Mars. Mons Olympus, the biggest volcano in the solar system, is 23 km high! (Figure 6). Not all basalt volcanoes have the gentle slopes characteristic of those in Hawaii. Many have the simple conical shape that one always associates with volcanoes. These are *composite cones*, constructed from

composite
cones

FIGURE 6 Olympus Mons. This 23 km high volcano on Mars is the biggest volcano in the solar system. The complex crater at its summit is about 70 km in diameter, and the cliffs around its base are about 4 km high. It is thought that the volcano is built up on a crust about 200 km thick. (NASA Mariner 9)



both lavas and pyroclastic rocks. They owe their elegant, sweeping profiles mostly to the fact that as the volcano is built upwards by material added through the vent at its centre, so erosion on its flanks removes material and spreads it out radially (Figure 7).



FIGURE 7 Two composite cones. (a) is a small cone, forming the volcano Vulcanello in the Mediterranean; the alternating layers of lava and ash of which it is made can be seen in the cliffs. (b) shows the summit area of Mt. Etna, over 3000 m high, the biggest volcano in Europe. The slopes on the left are covered with ash; lavas are present in the foreground behind the pylon, which formed part of a ski-lift destroyed by the eruption of 1971.

3.3 Constructive plate margins and the origin of basalts

In Units 6 and 7 you saw that basalts are erupted in enormous volumes from volcanoes along constructive plate margins. You also saw that the crust is very thin beneath the oceans—the mantle is only a few kilometres beneath the surface. It follows, therefore, that the basalts that have erupted from constructive plate margins have more than a little to do with the peridotites of the mantle. Table 3 gives chemical analyses of a typical oceanic basalt and a peridotite; look carefully at these data.

What are the main differences between the two rocks?

There is much more aluminium, calcium, sodium and potassium in the basalt than in the peridotite, and much less magnesium. Now, in Units 6 and 7, it was stated that the mid-ocean ridges are sited on top of ascending convection currents of molten mantle material.

What composition would the molten liquid have if the peridotite of the mantle were *completely* melted?

Obviously, it would have the *same* composition as the original peridotite.

If only *part* of the peridotite is melted, however, the results are quite different. If

you melt a simple, pure substance such as ice, H_2O , the first drops of the liquid are bound to be water also. But rocks are not simple, pure substances—they are composed of mixtures of minerals. When a peridotite is melted, the *first* drops of liquid are *not simply molten peridotite*.

A 'fresh' mantle peridotite consists mostly of the mineral olivine, with smaller amounts of pyroxene and a little plagioclase feldspar. Bearing in mind what you have read in Section 2.3, can you suggest what the composition of the first drops of liquid will be when peridotite is melted?

Since pyroxene crystallizes at a lower temperature than olivine, it will start to melt earlier, and thus the first liquid will tend to have a pyroxene composition. Similarly, since iron-rich minerals crystallize at lower temperatures than their magnesium-rich counterparts, the first liquid will tend to have more iron than magnesium.

Thus, the composition of the first liquid does not bear any *simple* relation to that of the solid rock. In the case of peridotite, the first drops of liquid have a composition that is best approximated by a mixture of pyroxene and plagioclase feldspar. The fact that partial melting yields liquids of different composition from the solid starting material is an extremely important one in geology, and helps to explain how it is that the Earth has a crust of the kind that we are familiar with, and how an enormous range of different rocks exists in different environments.

The significance of this discussion of the partial melting of peridotite is that it is generally thought that basalts—which consists mostly of a mixture of pyroxenes and plagioclase feldspar—are created at constructive plate margins by *partial melting of mantle peridotite*. The details of the melting process are complex, and need not concern us here. The important fact is that it is partial melting of mantle peridotite to form basalts that provides the source of all the material required to generate fresh oceanic crust as sea-floor spreading proceeds.

Look carefully at the basalt and peridotite analyses in Table 3. Which elements are concentrated in the basalt relative to the peridotite, and by how much?

Many of the elements are strikingly more abundant, but titanium, sodium and potassium are all roughly ten times as abundant.

Does this suggest anything to you about what proportion of the peridotite has to be melted in order to make basalt?

The answer to this question may not be immediately obvious. But think about it this way. If all the K_2O in the peridotite (0.03 per cent) goes into the basalt (0.26 per cent) there has been a concentration factor of $0.26/0.03 = 8.6$. In other words, the K_2O which was originally in the peridotite has been distributed into a mass of basalt which is $1/8.6$ of the mass of the original peridotite. This is approximately 11 per cent.

There are many other factors that have not been taken into account here, but calculations like these enable geologists to estimate the fraction of mantle peridotite that is melted in making basalts; in general, the fraction is less than 10 per cent.

3.4 Plutonic rocks; crystal fractionation and partial melting

The basalts that are erupted by volcanoes at constructive plate margins are only the visible surface manifestations of processes taking place under ground. *Plutonic rocks* are an important group of igneous rocks that originate as liquid magmas, like basalt, but are intruded or squeezed into other, older rocks. Plutonic magmas never reach the surface—Pluto was the mythological God of the Underworld, hence his name is well applied to rocks that never see the light of day. Plutonic rocks cool from their liquid state extremely slowly, since they are well insulated by their surrounding rocks. As explained on Audio-visual tape AC 90, this means that when they crystallize and solidify, their crystals have plenty of time to grow to large sizes.

It is clear that the centre of a large pluton will remain hot for an extremely long time—perhaps thousands of years—and, therefore, its centre will be very coarse grained. But what will happen at the *edges* of the pluton?

Clearly, where the pluton comes up against the surrounding rocks, it will be rapidly chilled, and a much finer grained rock results on crystallization. The degree of chilling, of course, depends on how cool the surrounding rocks are; at great depths in the Earth's crust, they may not be very cool at all. Because rocks are such poor conductors of heat, the thickness of the fine-grained *chilled margin* is never very great—it may be a centimetre or less in a metre-thick dyke, a metre or so in a 1 km diameter pluton.

chilled margin

There are thus three definite characteristics of plutonic rocks:

- 1 they cut across older rocks;
- 2 they are coarse grained, with chilled margins;
- 3 they are formed underground.

So far, we have discussed rocks of basaltic compositions only. But now consider what would happen if a large amount of basaltic magma, produced by partial melting of mantle peridotite, were intruded into a shallow part of the crust, say a couple of kilometres below the surface. The first thing that would happen, of course, is that a chilled margin would be produced all round the pluton, at the contact with the surrounding rocks. The chilled margin would have more or less the same composition as the basaltic liquid.

But now we have a large volume of hot basaltic magma, well insulated from the outside air. Crystallization of this large volume will continue over a very long period. The details of the crystallization process are complex, and we need not go into them here, but an important principle emerges from a brief examination of the process. Just as the first melt to be produced on partial melting does *not* have the same composition as that of the original rock, so the first crystals to form from a melt do *not* have the same composition as the melt itself. The process of crystallization therefore offers another way of making rocks with different compositions. Consider the data in Table 9.

TABLE 9 Chemical compositions of basalt and olivine

| | Basalt weight per cent | Olivine weight per cent |
|--------------------------------|---------------------------|----------------------------|
| SiO ₂ | 49.2 | 40.0 |
| TiO ₂ | 1.4 | |
| Al ₂ O ₃ | 15.8 | |
| Fe ₂ O ₃ | 2.2 | |
| FeO | 7.2 | 10.0 |
| MgO | 8.5 | 50.0 |
| CaO | 11.1 | |
| Na ₂ O | 2.7 | |
| K ₂ O | 0.3 | |
| Others | 1.6 | |
| | 100.0 | 100.0 |

Olivine is one of the commonest minerals in igneous rocks. It is important to us in the present context because when a basaltic magma begins to crystallize, the *first* crystals to form are olivine, as you saw in Section 2.3.

As you can see from Table 9, olivine contains silica (SiO₂), magnesium (as MgO) and iron (as FeO). When crystals of olivine form, these components are therefore removed from the melt.

What, therefore, is the effect of crystallizing olivine on the composition of the melt?

Clearly, the melt will be relatively depleted in the elements going into the olivine—iron and magnesium. Less obviously, the composition of the melt will be *enriched* in the elements not used, such as sodium and potassium. Less obviously still, the melt will be relatively enriched in silica because, although *some* is used in the olivine, olivine uses *more* of the other elements than it does silica.

These important relationships can be best understood by working through a particular case. Consider a mass of 100 tonnes of magma, from which ten tonnes of olivine have crystallized (we'll see what happens to the crystals in a minute). The ten tonnes of olivine contain four tonnes of silica (SiO_2), one tonne of iron oxide (FeO) and five of magnesium oxide (MgO). Thus, these ingredients are subtracted from the liquid, and the 90 tonnes of liquid remaining therefore contain the weights of the other elements shown in Table 10.

If you compare analysis (b) of the liquid in Table 10 with that of the basalt in Table 9, you will see that there are some subtle differences, which are illustrated in the histogram of Figure 8. You might be thinking at this point that subtle differences in columns of figures are only likely to be interesting to an accountant. They are, however, vitally important to geologists, because they reveal that two quite different rocks can be made, simply by subtracting crystals of olivine from a basalt magma. This process is known as *fractional crystallization*.

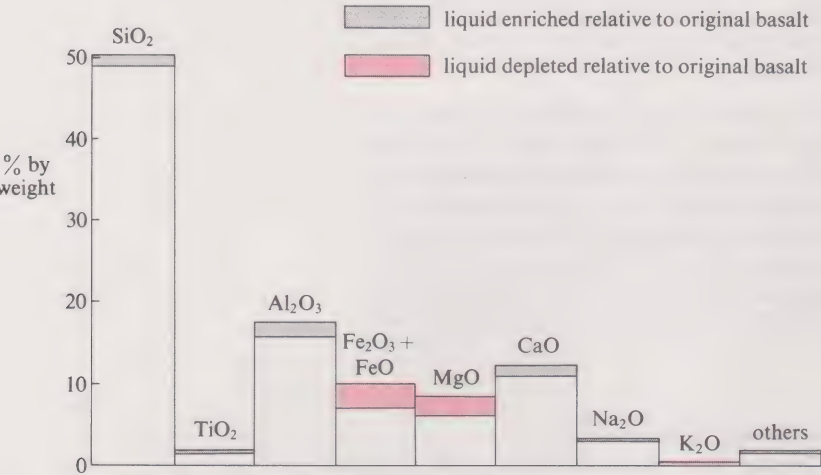


TABLE 10 Proportions of ingredients remaining in 90 tonnes of olivine-depleted liquid basalt: (a) in kilograms, and (b) as weight per cent

| | (a) | (b) |
|--------------------------------------|-------|-------|
| SiO_2 | 45.2 | 50.3 |
| TiO_2 | 1.4 | 1.6 |
| Al_2O_3 | 15.8 | 17.5 |
| $\text{Fe}_2\text{O}_3 + \text{FeO}$ | 8.4 | 9.3 |
| MgO | 3.5 | 3.9 |
| CaO | 11.1 | 12.3 |
| Na_2O | 2.7 | 3.0 |
| K_2O | 0.3 | 0.3 |
| Others | 1.6 | 1.8 |
| | 90 kg | 100.0 |

FIGURE 8 Histogram showing the differences in composition brought about by subtracting olivine from a basalt magma

In the case that we are considering, the olivine crystals that form in the magma may grow into substantial lumps that are denser than the surrounding magma, and hence sink to the bottom of the pluton. They will accumulate at the bottom of the pluton to form rocks known, not surprisingly, as *cumulates*. Since the cumulate consists of olivine crystals, the rock will have the composition of olivine. Cumulate rocks consisting mostly of olivine are common; more rarely, the cumulate minerals are oxides of iron, titanium and chromium, and these are of major economic importance.

If the liquid left behind when olivine crystals had been removed could be somehow abruptly chilled, it would have the composition represented in column (b) of Table 10, which is effectively basalt minus olivine. But it is much more likely that crystallization will continue, in which case olivine will be joined by other minerals such as pyroxenes, with different compositions following the sequence outlined in Table 4. These minerals will subtract other elements from the liquid, so the composition of the liquid will steadily depart from its original nature. When almost the whole of the pluton has crystallized, the last small liquid fraction may have a composition vastly different from the original, and may solidify to form a rock totally unlike basalt in every way: granite. (It is important to realize that *not* all granites are formed from basalts by fractional crystallization.)

ITQ 4 Look at the analyses of peridotite, basalt and granite in Table 3. Can you see any consistent trends in the proportions of the elements in each? And can you suggest what significance this might have in terms of crystal fractionation?

It is implicit from the answer to ITQ 4 that there is a continuous spectrum of rock compositions between peridotite and granite, characterized in particular by increasing SiO_2 content. Fractional crystallization and partial melting are two ways in which rocks of different compositions can be made, but it is important to grasp that *both* these processes operate in the same direction, towards increasing silica content or, in other words, towards rocks of increasingly granitic nature. Thus,

whereas a peridotite can be partially melted to make a basalt, a peridotite liquid can also be fractionally crystallized to make basaltic and granitic rocks.

In both partial melting and crystal fractionation, there are some rocks that are 'left behind'. In the case of partial melting of mantle peridotite, what remains is a large volume of 'depleted' peridotite, from which the basalt has been squeezed out. This remains in the mantle, and is rarely seen at the surface. Similarly, crystal fractionation produces cumulates as well as silica-enriched liquids. The cumulates, necessarily, tend to be formed at deep levels, and are therefore not as frequently encountered as the rocks made from the remaining liquids. It is these that we deal with in the next two Sections.

One final point. We have been discussing here what happens to the compositions of basaltic liquids if crystals are *removed* from the liquid, for example, by the settling out of crystals. If crystals *cannot* be removed—if, for example, the liquid is too stiff or viscous to allow them to settle—then, clearly, the composition of the rock that is ultimately formed will be the same as that of the basalt. Such rocks, of course, will be coarse-grained basalts, or *gabbros*. These exist in huge quantities beneath the basalts of the oceanic crust.

3.5 Destructive plate margins and the origin of andesites

You were introduced in Section 5.3 of Units 6 and 7 to the volcanic processes that accompany the descent of oceanic crust into the mantle at a destructive plate margin. In this Section, we are going to look more closely at the processes that take place there. Consider first what happens in the simplest case—where one slab of oceanic crust is forced down beneath another.

Can you recall what kind of surface features characterize such a subduction zone?

It occurs at plate margins along which *island arcs* have been constructed; the significance of these island arcs is that they are located in the oceans, so there is no question of there being any continental crust beneath them. Now, when a slab of oceanic crust is forced down along a subduction zone, there is a great deal of friction beneath the descending and overriding plates. As you have already seen, this friction is manifested in the form of intense seismic activity, and in the liberation of heat. The descending slab of oceanic crust is also thrust deep down into the hot mantle, and is heated by conduction. The heat produced by friction and by conduction ultimately leads to the generation of magmas erupted by the volcanoes along island arcs.

So sufficient heat is generated to melt some rocks. But what is it that melts?

There are two possibilities: either some of the rocks of the oceanic plate melt *or* those of the overriding mantle against which it is grinding. Both, of course, might also melt to some degree.

Whichever rock it is that melts, would you expect the magma to have the *same* composition as the melted rock?

No, you should recall from Section 3.2 that when rocks are melted, the first melts that are produced do *not* have the same composition as the melting rock. Thus partial melting of the basaltic rocks of the oceanic plate would *not* produce a new generation basalt magma. Since the rocks of the overriding plate are mantle peridotites, you might expect that partial melting of these would produce basalt magmas, just as they do beneath the ocean ridges.

Unfortunately, life is not quite so simple. Beneath *constructive* margins, new mantle comes towards the surface and basaltic magma is extracted from it. The depleted mantle then moves on and diverges outwards, and the next bit of mantle in the conveyor system comes up and goes through the same process. On a gross scale, one bit of mantle has the same composition as the next, so the new oceanic crust produced from it will have the same bulk composition.

At *destructive* margins, by contrast, although the down-going plate moves continually, the overlying plate stays in the same position with respect to the subduc-

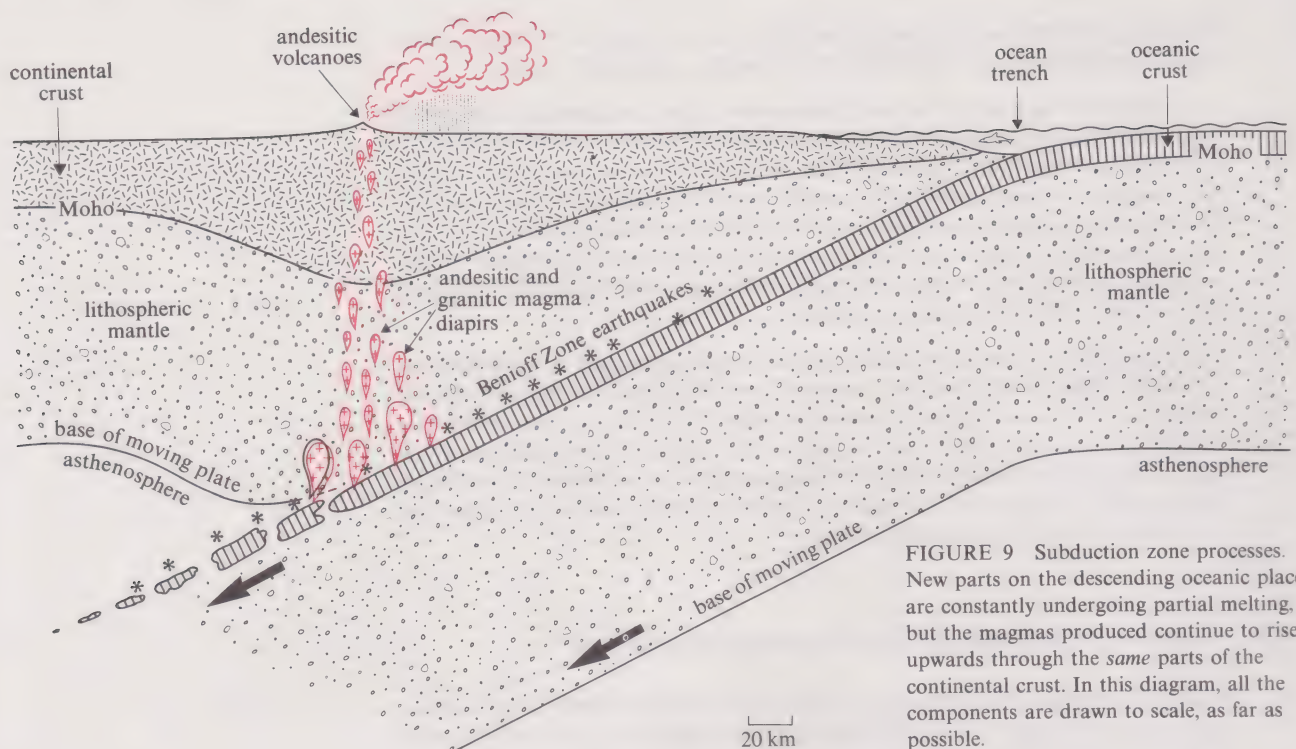


FIGURE 9 Subduction zone processes. New parts on the descending oceanic plate are constantly undergoing partial melting, but the magmas produced continue to rise upwards through the *same* parts of the continental crust. In this diagram, all the components are drawn to scale, as far as possible.

tion zone beneath it (Figure 9). This being so, magma is continuously produced at the same site, relative to the overriding plate. As time passes, more and more melts will pass upwards through the *same* zone in the overlying plate, which will be progressively heated up and will change in composition as magma is either added to or subtracted from it. In these circumstances, it follows that the first magma will differ markedly from those developed later on. In areas where we know that subduction is new and volcanic activity commenced recently, the first lavas are basaltic, similar in composition to those of the oceanic crust. The later products are quite different from these primitive basalts, and are much more siliceous.

These siliceous rocks, which are extremely characteristic of destructive plate margin volcanoes, are known as *andesites*. The name, of course, is derived from the Andes, where such rocks are very abundant. They are generally lighter in colour than basalts and lower in density.

andesites

What do these two facts suggest to you about the minerals present in andesites?

The lighter colour and lower density indicate that they contain *fewer* iron- and magnesium-rich minerals.

Andesites, in fact, are a sort of halfway stage between basalts and granites in almost every respect, as you might expect from an examination of their chemical compositions (Table 3).

There is one vitally important difference between andesites and basalts. Whereas basalts are created from mantle rocks at constructive plate margins and return to the mantle at destructive plate margins, the manufacture of andesites is a one-way process; andesites never return to the mantle. Thus, when a volcano at a subduction zone erupts, it not only provides a firework display of lavas and ashes but also yields an *irreversible addition to the continental crust*.

In Unit 4 it was stated that, broadly speaking, the continental crust is composed of a low-density igneous rock between granite and basalt in composition, with a veneer of sediments. The rock intermediate between granite and basalt is, of course, andesite, and it will be argued in this Section and the next that the continents themselves originated through andesitic volcanism at destructive plate margins. (The terminology is a little confusing here: oceanic crust is created at *constructive* plate margins, but continental crust is created at *destructive* plate margins!)

Consider for a moment a hypothetical Earth—maybe even the primordial Earth—in which there are *no* continents, but there is a hot, mobile mantle in which convection currents are active. Plate tectonic processes of a sort will operate—they do not require the presence of continents—and, in particular, there will be a number of destructive plate margins at which andesites are being erupted at island arc volcanoes. With the passage of time, some arcs will be rapidly eroded, and may even disappear below sea-level. But the material will *not* disappear into the mantle—for one thing, its density is too low.

Along the site of the inactive island arc, then, there will be a long, narrow belt of rocks of andesitic composition, mostly sediments, but with some of the original lavas intact. This inactive belt will drift around passively on the surface of the Earth for tens or hundreds of millions of years. Eventually, however, plate tectonic processes are bound to lead to its colliding with *another* arc, possibly an active one (Figure 10).

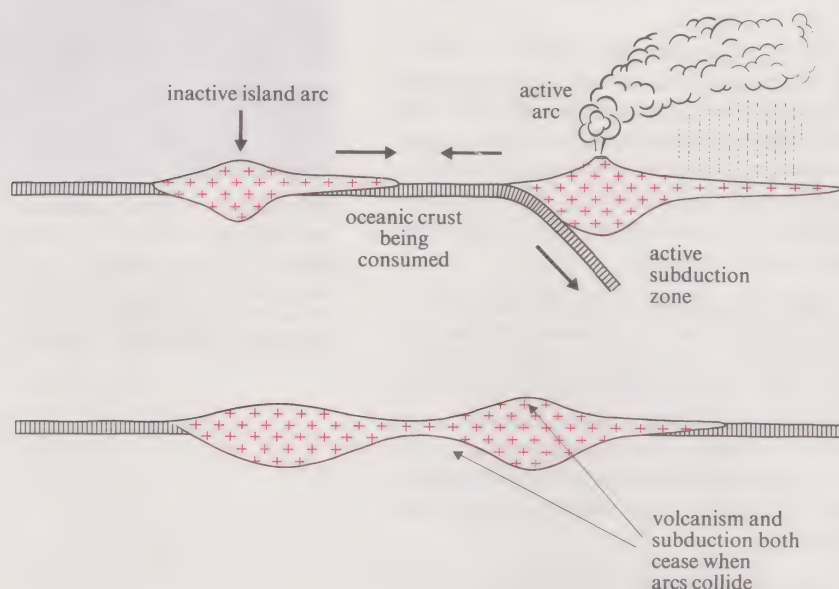


FIGURE 10 Cartoon illustrating the accretion of island arcs to form a mini continent.

Because of its low density, the ancient island arc *cannot* be forced down into the mantle beneath the new arc—it would be a bit like trying to push a beachball under water. The net result is that the old island arc clangs up against the new one, subduction ceases, and the two become welded together to form a ‘mini continent’. With the continuation of plate tectonic processes, it is easy to see that as more and more island arcs are accreted together, the size of the accreted mass will increase until it reaches continental dimensions.

island arc accretion

3.6 Granites and the origin of continents

You will already be familiar with the granite specimen S1 in your Home Kit and, hopefully, will also have determined its density. Your specimen is a typical plutonic igneous rock; it is coarsely crystalline, the crystals are tightly interlocked with one another, and the different minerals present are easily recognizable.

Whereas basalts are erupted at temperatures of over 1 100 °C and flow freely like ordinary liquids, granites melt at *much* lower temperatures and the viscosity of granite melts is much higher. It is thought, in fact, that a granite magma is more like a pasty mush than a free-flowing liquid, hence, when crystals form in it, they do not settle out but grow larger *in situ*, eventually interlocking with other crystals.

When you determined the density of your granite specimen, you should have found that it was much less dense than any of the other igneous rocks. Consider a granite magma of density $2.4 \times 10^3 \text{ kg m}^{-3}$, which is formed at the base of continental crust with density $2.9 \times 10^3 \text{ kg m}^{-3}$. Since it is less dense than the surrounding rocks, it will try to ‘float’ up to the surface. Granitic magmas, therefore, have a powerful upward urge, and this enables them to rise great distances through the crust.

If such a rising granite body enters a region of rocks of lower density, it will cease to rise and, similarly, if it cools and solidifies, it will grind to a halt. Until either of

these two things happens, granites will rise through the crust, pushing aside the surrounding rocks. In some cases, they come extremely close to the surface, a matter of one or two kilometres. In other cases, they come so close to the surface that a conduit is blasted through to the open air, and magma makes its way to the surface. On doing so, of course, the granite ceases to be a *plutonic* rock and becomes a *volcanic* one (Figure 11).

Granites are found in huge volumes at destructive plate margins, especially in areas such as the Andes where the continental crust is very thick. But granites form only a few per cent of the dominantly 'andesitic' continental crust. How then do they form? To answer this, we need to look more closely at the igneous processes taking place at destructive plate margins. We have said that, in the simplest case of an island arc, basaltic and andesitic magmas are produced by partial melting of mantle material. Granites are formed by carrying this process a stage further; in a sense, they are distilled or refined andesites, the results of further partial melting.

diapirs What seems to happen is that in its early stages, the overriding plate at a subduction zone is relatively cold and brittle. Early formed andesite magmas aggregate into blobs or *diapirs*, which then rise up into the upper crust. Initially, the melts will not rise very far before they solidify. However, the process is continuous, and as more and more diapirs rise, so the base of the overriding plate becomes hotter and hotter. Eventually, partial melting of some of early formed rocks takes place, to give more siliceous granite magmas, which then rise to higher levels in the crust before solidifying to form coarse-grained plutonic rocks.

In this way, at an island arc where initially nothing but oceanic crustal rocks are present, increasingly siliceous crust is generated, which becomes progressively more continental in character. This process seems to take place relatively quickly because, in areas where island arcs are no more than 20 million years old, continental crust up to 30 km thick has been identified by seismic studies. Indeed, there is growing evidence suggests that much of the crust in continents like Africa and Asia was formed by processes of island arc granitization over the last 1 000 million years. It is also possible that most of the oldest continental rocks, more than 2 600 million years old, were formed in this way.

Once an island arc has developed a siliceous foundation, it is well on its way to becoming a 'mini continent'. If subduction zone activity should be renewed beneath an old island arc, it can only result in yet more siliceous magmas, and in the steady thickening of the continental crust.

Summary of Sections 3.5 and 3.6

The net result of all the processes taking place at subduction zones is that throughout geological time there has been a slow and steady conversion of mantle peridotite to andesitic continental crust. This is an extremely important result, so it is worth summarizing the most important steps:

- 1 Partial melting of peridotite at destructive plate margins forms basaltic and andesitic magmas.
- 2 Andesites accumulate irreversibly to form island arcs.
- 3 Prolonged magmatism beneath island arcs leads to remelting and progressively more siliceous magmas, which eventually reach granitic composition.
- 4 Island arcs accrete to form 'mini continents'.
- 5 Renewed destructive plate margin adds further siliceous magmas, thus increasing the thickness of the continental crust and enlarging the 'mini continents'

Objectives of Section 3

When you have finished reading this Section you should be able to:

- (a) Describe the possible sources and locations of the Earth's internal heat. (SAQ 3)
- (b) Explain systematically, and solve simple problems based upon, the ways in which the compositions of igneous rocks can be changed, and new rocks formed, by partial melting and fractional crystallization. (SAQs 4, 7, 8–10)



FIGURE 11 The elliptical structure at the centre of this photograph is 40 km in diameter; it was produced by the approach of a granitic pluton to very near the surface of the Earth. (LANDSAT satellite photograph, north-west Argentina)

- (c) Summarize the main differences between volcanic and plutonic rocks. (SAQ 3)
- (d) Predict the heights to which volcanoes will grow in given circumstances. (SAQ 6)
- (e) Explain how continents are formed by destructive plate margin processes. (SAQs 7 and 10)

SAQ 3 Which of the statements below correctly describe the source of the Earth's internal heat? Give a brief statement of your reasons.

- A Most of the Earth's heat comes from radioactive decay of elements in the Earth's crust.
- B The rate of radiogenic heat production within the Earth must always have been the same as it is today.
- C Lead in the core of the Earth could contribute towards its internal heat.
- D Not all the Earth's heat need be of internal origin. Some must come from the Sun.
- E The Earth may retain within it heat generated during the formation of the Earth 4.6 billion years ago.
- F The small quantities of radioactive elements within the mantle are inadequate to account for the convection cells moving within it.
- G Heat may be generated through frictional effects due to the movement of convection cells within the mantle.

SAQ 4 Which elements would be relatively depleted in the residual liquid left when plagioclase crystals consisting of 100 per cent pure $\text{CaAl}_2\text{Si}_2\text{O}_8$ are removed from an andesite magma? And which elements would show the greatest relative depletion? Explain your answer. (You may assume that the composition of the plagioclase is: SiO_2 57%; Al_2O_3 30%; CaO 13%).

SAQ 5 Which of the following terms or statements apply *only* to volcanic rocks? (Some are irrelevant or confusing.)

- A Cross cutting relationships
- B Dykes
- C Older than surrounding rocks
- D Coarse-grained margins
- E Fine-grained margins
- F Coarse-grained all through
- G Fine-grained all through
- H Resting on top of younger rocks
- I Resting on top of older rocks
- J Formed by partial melting
- K Formed by fractional crystallization
- L Formed underground

SAQ 6 Granite magmas have a density of around $2.6 \times 10^3 \text{ kg m}^{-3}$. How high could a volcano erupting rocks of granitic composition grow above a lithospheric plate of density $3.3 \times 10^3 \text{ kg m}^{-3}$ and 70 km thick?

SAQ 7 There are 50 000 km of active destructive plate margins in the world. Assuming that each kilometre contributes 25 km^3 of magma to the surface per million years, and that 80 per cent of this is derived from the mantle, how much *new* continental crust is generated each year?

SAQ 8 Where do you suppose the other 20 per cent of the magma (see SAQ 7) not derived from the mantle comes from?

SAQ 9 The vast majority of granites in the world are formed at destructive plate margins by partial melting of mantle and older continental rocks. By what other mechanism could small volumes of granitic magma be made?

SAQ 10 Fill in the missing words in the following paragraph from those in the key.

Granites may be formed in small volumes at plate margins by extreme crystallization of magmas, but they are formed in much larger volumes at plate margins, where progressive episodes of magmatism lead to the evolution of progressively more liquids. Heat carried up into the crust may also cause partial in the upper crust, thus contributing further to the volume of granitic magma.

KEY

| | |
|-------------|--------------|
| basaltic | constructive |
| melting | siliceous |
| destructive | fractional |

4 Sediments and sedimentary rocks

Study comment Before reading this Section, you should complete the Home Experiment for Unit 26, if you have not already done so. The results of this experiment are particularly relevant to this Section.

4.1 Introduction

In going on to look at the ways in which sediments and sedimentary rocks are formed, we are moving on to the next stage in the rock cycle. This is the stage that takes over once igneous-rock-forming processes are complete and the igneous rocks are exposed at the Earth's surface, either because they were erupted initially as lavas or because the rocks into which they were intruded have been worn away. Such igneous rocks are exposed to a totally different environment at the surface of the Earth from the one in which their constituent silicate minerals originally crystallized—temperature and pressure are both much lower, and there is an abundant supply of oxygen-rich water.

In response to these environmental differences, the silicate minerals of the igneous rocks undergo changes that may lead to their total or partial breakdown and, consequently, to the destruction of the rocks themselves. These breakdown processes are known collectively as *weathering* and it is the products of weathering that eventually accumulate to form sedimentary rocks. This is one part of the rock cycle that you can see for yourself in all parts of the country.

weathering

4.2 Physical and chemical weathering of rocks

Two distinct types of weathering can readily be distinguished: *physical weathering* (or mechanical disaggregation) and *chemical weathering* (or chemical decomposition). It is combinations of these two processes that lead to the formation of a wide range of sedimentary rocks.

physical weathering
chemical weathering

One of the best-known cases of physical weathering, causing the break up of large masses of rock into fragments, results from the freezing of rain-water in cracks and fissures. Unlike most other substances, water *expands* on freezing with a 9.2 per cent increase in volume and the resulting pressure increase on the sides of the fissures causes them to split further apart. Large blocks of rocks fall under gravity to the front of rock faces, where they are shattered into smaller fragments which accumulate as scree slopes. As more material is added to the scree slope, this shifts to maintain stability and so the fragments move progressively downslope, shattering even further as the result of inter-particle collision. Eventually they will be removed from the foot of the scree slope altogether by an agent of transportation to be deposited later as sediment (see Section 4.3). The splitting of rocks by *frost shattering* is responsible for the broken slabby boulders shown in Figure 12.

frost shattering



FIGURE 12 Frost-shattered boulders. The cracks and crevices in which water collects and freezes can clearly be seen in the boulders and various stages in the break-up of boulders are visible. (Institute of Geological Sciences photo)

Exposed rock faces and weathered scree fragments are exposed to rain-water run-off and soon undergo chemical weathering. This attack is enhanced by the presence of dissolved atmospheric gases such as carbon dioxide, which make the water slightly acidic. Chemical reactions may take place between the rock minerals and water, sometimes dissolving them completely, or else producing new mineral types. Chemical weathering takes place more effectively when *physical* weathering is well advanced, as the progressive fragmentation of large rock masses increases the total surface area exposed.

You will recall from Section 2.3 that the silicate minerals of igneous rocks crystallize in a sequence controlled by their temperature of formation. Each of these minerals varies in its susceptibility to chemical attack, a property that is described as the *relative resistance* of the mineral.

Look at Table 11, which lists common minerals in order of their relative resistance. How does this compare with the order of crystallization shown in Table 4?

The first-formed, high-temperature silicate minerals are the first ones to break down, whereas the last-formed, low-temperature minerals are the most stable.

TABLE 11 The relative resistance of silicate minerals to weathering

| | Iron- and magnesium-rich minerals | Iron- and magnesium-poor minerals |
|---|-----------------------------------|---|
| Increasing resistance to chemical weathering ↓ | olivines | calcium- and sodium-plagioclase feldspars |
| | pyroxenes | K-feldspar |
| | amphiboles | muscovite-mica |
| | biotite mica | quartz |
| | | |

In fact, water has virtually no chemical effect on quartz at all. This sequence of silicate mineral stability is not coincidental, but is directly related to the chemical compositions and structures of the minerals themselves. The covalent silicon-oxygen bonds are far stronger in silicate minerals than the ionic bonds between the silicate anions and any other cations. Therefore, the minerals with the *highest* silicon-to-oxygen ratios are less easily broken down than those with lower ratios. Silicates with high silicon-to-oxygen ratios (Si : O) are those with complex structures and much oxygen sharing, which form at low temperatures, such as quartz (Si : O = 0.5). Silicates with low Si : O ratios are those with simple structures, which form at high temperatures, and have little or no oxygen sharing, such as the olivines (Mg, Fe)SiO₄, where Si : O = 0.25. Minerals with low Si : O ratios also have high cation contents and it is the leaching out of these cations that leads to chemical decomposition.

relative resistance

4.3 The products of chemical weathering

When an igneous rock such as a basalt or a granite is chemically weathered, various products are formed, which fall into three main categories:

- (i) Substances that are soluble in water and are carried away in solution by rain-water and rivers, eventually entering the sea.
- (ii) New minerals formed when soluble substances are leached out of existing minerals and their atomic structures partially collapse.
- (iii) Resistant minerals, which are not chemically attacked at all.

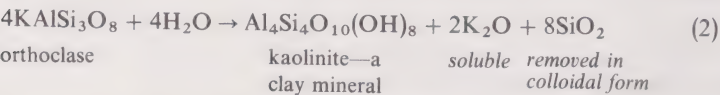
The new minerals and the resistant minerals are known as *residual minerals*. The most common of the resistant minerals is quartz but, if chemical weathering is not too prolonged, feldspar and some micas may also resist chemical attack. Quartz, of course, is found in huge volumes as a residual mineral in sand grains in beaches all over the world. Paradoxically, perhaps, many of the finest dazzling white beaches that holiday makers throng to are made of grains of calcium carbonate, which is not at all resistant to chemical attack. By far the most highly prized residual mineral, though, is *diamond*. Diamonds have a complex origin, and are formed in igneous rocks at great depths and pressures within the Earth. When the parental rocks are exposed at the surface, they are eroded and the diamonds are washed out, and may be transported hundreds of miles by rivers. The diamonds found in river gravels show no trace of their ordeal and form a major source of the world's diamonds.

residual minerals

The breakdown of iron- and magnesium-rich minerals such as olivines usually leaves a residue of insoluble iron oxides. When the soluble cations are leached out of the more complex silicates such as feldspar, the collapsed atomic structure usually takes in water to form a flakey, hydrated *clay mineral*.

clay mineral

We can show some of the products formed from the chemical weathering of silicate minerals by the reactions of Mg-rich olivine and orthoclase feldspar (K-feldspar) with water*:



These equations are somewhat simplified, as natural rain and river waters also contain dissolved atmospheric CO₂ (see Section 4.2.2). There is more than just academic interest in the breakdown of feldspar to clay mineral in equation 2: in Cornwall, the feldspars in vast quantities of granite have been broken down to form valuable deposits of *china clay*, which have been worked commercially for decades. China clay is used extensively to whiten paper, like the pages you are reading!

4.4 The transportation of weathered material

You have already seen that the soluble products of chemical weathering are removed by rain-water and rivers. However, the residual products of chemical weathering and the larger fragments produced by physical weathering are also removed by various transporting media. The removal of weathered material is usually referred to as *erosion* and the combined processes of weathering and erosion that gradually wear down mountainous landscapes are termed *denudation*.

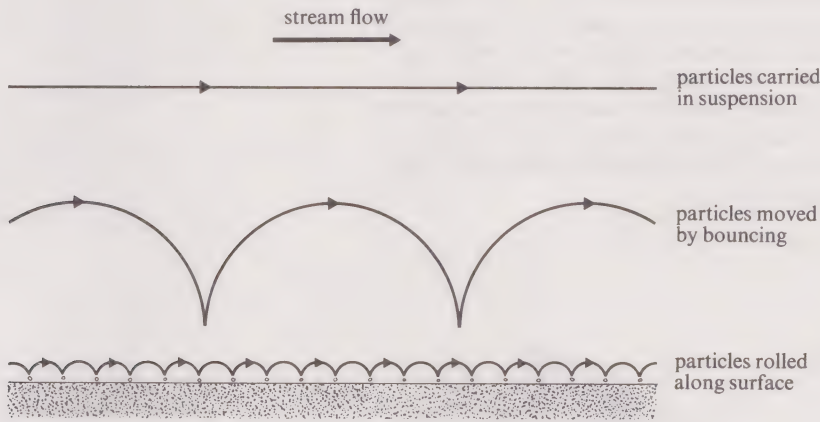
erosion

denudation

In addition to water, wind and ice can also be important transporting media, although fortunately the latter two are not significant in Britain at the present day. However, as you saw in TV programme 6, there is evidence in our geological record to show that there have been periods of time when Britain was subjected to wind erosion in desert conditions, and scoured by glaciers and ice sheets. As it is water movement that concerns us in Britain today, it is this that we shall concentrate on.

*You are not expected to remember these formulae.

The transportation of rock fragments and mineral grains by water involves three types of movement (Figure 13). Some particles are *rolled* along the bed of streams, others roll until they collide with each other or are lifted up temporarily by eddy currents and carried in the water for short distances until they sink back to the ground again. This form of movement in a series of erratic jumps is called *saltation*. The rolled and saltated particles form the transported *bed-load*. Finer grained particles are carried within the water itself and form the *suspended load*.



rolling

saltation
bed-load
suspended load

FIGURE 13 Modes of sediment transportation in water, showing rolling, saltation, and suspension.

Clearly, grain size plays an important part in determining how sedimentary particles are transported. In your Home Experiment you will have already investigated the relationship between the size of a grain and the rate at which it falls through a fluid. The relationship, $v \propto r^n$, is known as *Stokes' law*.

Stokes' law

The relationship also assumes that the particles are perfect spheres. In natural environments, of course, this is a most unlikely situation. You have already seen in the Audio-visual sequence that particles may vary from completely angular to nearly rounded in shape. The less well rounded a particle is, the more resistance it offers as it falls through the water.

In the Home Experiment you used a stationary body of water and you found that small grains displaced the water smoothly whereas larger ones created turbulence. In natural systems, both particles *and* water may be moving. The same rules apply, however. Whenever flow past a particle is laminar, the particle falls according to the relationship $v = kr^2$. If there are random fluctuations in the velocity and direction of movement of the flowing medium, in other words, if the flow is turbulent, the relationship $v = kr^{1/2}$ applies.

Are large particles more or less likely to be involved with turbulent flow than small ones?

Clearly, the smaller the particle, the *less* likely is turbulent flow. The velocity and turbulence of flowing water is a measure of the *energy of the transportation environment* and, as grain size and velocity are related, the energy of the environment is the most important factor governing the transportation of particulate matter. It is the energy of the environment that determines whether a given grain size is transported or not. Water flowing with constant velocity will transport all particles up to a given maximum size; if the velocity decreases, the largest particles will settle out of suspension and be *deposited*.

energy of environment

Changes in water-current velocities are largely responsible for the natural selection and deposition of certain grain sizes in sediments, or the *sorting* of a sediment. Sorting is a quantitative term and is a measure of the size-frequency distribution of grain sizes in a sediment. Fast-flowing water in a high-energy environment carries all but the coarsest material, which is deposited; so coarse-grained sediments are characteristic of high-energy environments. Similarly, fine-grained sediments are characteristic of low-energy environments.

sorting

Can you remember from the Audio-visual sequence for Unit 4 (tape AC 90), where some high-energy environments may be found?

On storm beaches and in river channels during flood conditions.

The relationship between water velocity and grain size deposited is shown by the lower line in Figure 14. The results of your Home Experiment should fall close to this line.

ITQ 5 From Figure 14, what velocity is required for the deposition of fine sand grains 0.125 mm in diameter?

From Figure 14 you can infer that to deposit clay particles, which are less than 0.004 mm in diameter, current velocities of much less than 0.01 m s^{-1} are required. The slightest water movements will, therefore, maintain clays in suspension, so clay deposits are diagnostic of very low-energy environments, where water movement is negligible. The settling velocities of clay minerals are also reduced by their flakey shape, which offers considerable resistance to the water as they settle. This is why clays are commonly found in bodies of almost still water such as lakes and lagoons. They also occur in river estuaries, where deposition takes place during the period of slack waters accompanying high tide.

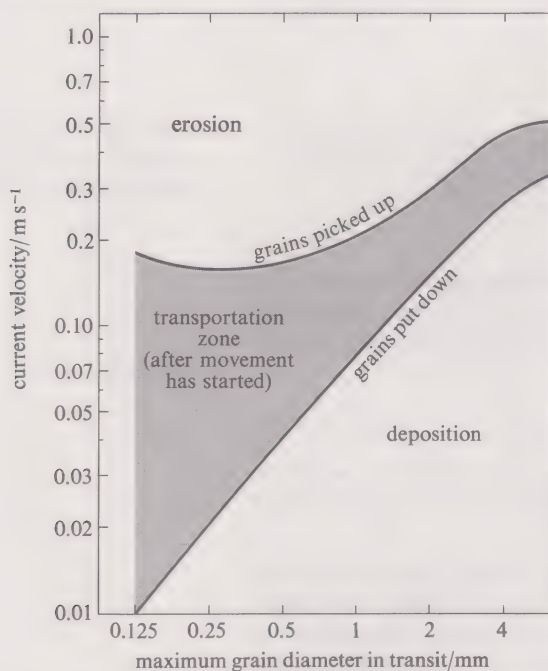


FIGURE 14 Graph relating size of sediment grains to the velocity of a water current required to pick them up (upper curve) and deposit them (lower curve). The zone between the two lines shows the field in which sediment grains can be carried in suspension by the current, once they have been picked up.

Zonation of sediment types often develops as a response to varying energy conditions. Rocky coastlines, with large boulders and pebbles strewn around the headlands, are present where wave action is concentrated, whereas sandy beaches (sometimes backed by pebble storm beaches) form in sheltered bays where wave action is more gentle. The constant wave action winnows out the finer sands and clays and maintains them in suspension to be carried up river estuaries and deposited on the incoming tide.

Sediments consisting predominantly of one grain size are said to be *well sorted*, whereas sediments containing a wide variety of grain sizes, deposited perhaps as the result of a large decrease in the energy of the environment, are said to be *poorly sorted*.

ITQ 6 Would you expect beach sands in sheltered bays to be well sorted or poorly sorted? Give reasons for your answer.

Sediment zonation can also be recognized in large, well-developed river basins. Near the source of the river, where its bed is steep and water flow is fast and turbulent, only large boulders and pebbles are found. As the gradient and velocity decrease down river, first coarse and then progressively finer sands are deposited. The clays collect on the flat flood-plains close to the estuary, or as mud-flats within the estuary.

Will these zones always be found in the same places in a river basin? If not, why not?

No, they will vary considerably as the volume and velocity of the river water fluctuates in response to changes in rainfall.

Thus, if you were to dig down through successive sediment layers in a river channel, you would find fine sands alternating with coarse sands and even gravels, reflecting alternate drier, wetter and flood periods.

4.5 The erosion of deposited sediment

Once a sedimentary particle has been deposited, that is not necessarily the end of its story. An increase in current velocity may cause the particle to be picked up again, or eroded. For this to occur, there must also be turbulent flow in the water creating an upwards component of motion. The upper curve on Figure 14 shows the current velocity required to pick up grains of a given size.

ITQ 7 Compare the current velocity required to erode a sand grain of 0.125 mm diameter with that required to deposit the same grain. How do they differ?

The forces required to lift and transport a deposited grain from the underlying surface must be sufficient to overcome the weight of the grain immersed in water, which acts as a vertically downwards force, and the frictional force between the grain and the underlying surface, which prevents the grain from sliding. As you can see from Figure 14, a considerable drop in velocity may be required to deposit the grain again once it has been eroded. The shaded area in Figure 14 shows the range of current velocities within which material of a given grain size will be transported in suspension.

Once deposited, clay particles require very substantial current velocity increases to erode them. This is because clay minerals have very strong cohesive properties, which causes them to adhere to each other firmly. The cohesive properties of clays are responsible for the rapid build up of estuarine mud-flats and explain why so much hard work is required to dig a clay garden soil! These cohesive properties are responsible for the marked turn-up of the top curve on the left part of Figure 14.

cohesion in clays

Particles that have been picked up and transported are subjected to mechanical abrasion, and undergo changes in shape. It is fairly obvious that the longer particles are in motion, the more eroded they will become, as edges and corners are worn away. It is much less obvious that *large* particles are much more rapidly rounded than small ones; this is partly because large particles have more momentum than smaller ones and hence can do more damage to each other when they collide. The upshot of this is that rounded particles are much more common in coarse-grained sediments deposited in a high energy environment than in fine-grained sediment deposited in a lower-energy environment. This is something you can easily confirm for yourself on a beach that has both pebbly and sandy parts. The pebbles are almost always smoothly rounded, but the sand grains are much more angular. (You may need a magnifying glass to see them.)

4.6 The transportation and deposition of the soluble products of chemical weathering

The most abundant cations released from silicate minerals by chemical weathering are Ca^{2+} , Na^+ , K^+ and Mg^{2+} . These, along with silica, are carried away in rivers, eventually reaching the sea. You might imagine that as weathered material in solution has been added to the oceans over a period spanning many millions of years, the sea should be becoming progressively more saline. This is not the case, because the dissolved salts are removed at the same rate as they are added, which means that the composition of sea-water has remained reasonably constant over at least the past 10^9 years. Two major ways in which dissolved salts are removed are by (i) the action of marine organisms and (ii) by direct chemical precipitation.

(i) *The action of marine organisms*

Many marine animals build shells composed of calcium carbonate, and they extract both Ca^{2+} and the anion HCO_3^- from sea-water to do this. The shell material is in the form of *calcite* (CaCO_3) and, after the deaths of the organisms, their calcareous remains may accumulate on the sea-floor. If they are present in sufficient abundance, they may become cemented with muds containing chemically precipitated calcite to form *limestone*. Many limestones are formed on the continental shelf regions, where there is an abundance of marine life. Evidence for this marine life is found in fossils which are often large enough to be seen clearly in a hand specimen. If you examine your Home Kit specimen S7 carefully, you should be able to see fossil remains in it. These are remains of shelly marine organisms, mostly molluscs. Ornamental slabs of polished limestone used in buildings often display such fossils spectacularly well, one example is the so-called 'Purbeck marble', used in many British cathedrals and churches.

calcite

limestones

Some calcareous sediments form in the deep ocean basins directly from the accumulation of calcareous microfossils, far too small to be seen with the naked eye (Figure 15). The unconsolidated sediments are called *calcareous oozes*. They eventually become compacted and cemented to form fine-grained chalks and limestones. Limestones exposed at the Earth's surface are themselves very susceptible to chemical weathering, and you have already seen in Unit 14, Section 5.5, how carbonated rain-water and river water is responsible for the solution of limestone.

ooze

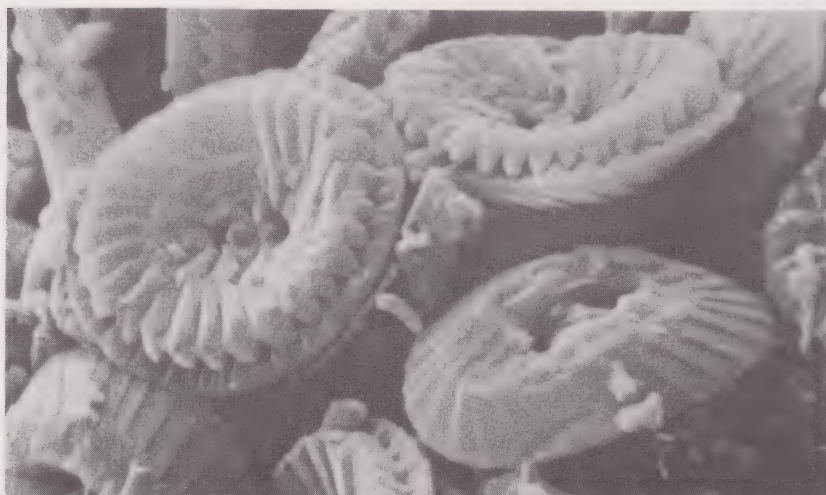
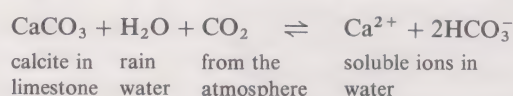


FIGURE 15 Calcareous microfossils. This is a scanning electron microscope photograph; the individual fossils are far too small to be seen with the naked eye. Some rocks, such as the chalk in the White Cliffs of Dover, are made up entirely of microfossils like these. (Institute of Geological Sciences photo)

You may remember that the chemical reaction that takes place is:



The weathering of limestones is a major source of HCO_3^- in river waters, which explains why water in limestone regions is 'hard'.

(ii) *Direct chemical precipitation*

Sediments formed in this way are comparatively rare, but some have great economic importance. The best known are *salt deposits* or *evaporites*, which form in isolated inland seas or in narrow, newly formed ocean basins where circulation with other masses of sea-water does not take place. Here, the evaporation of the water may exceed the input by rivers, so that the water becomes saturated with respect to ions such as Na^+ , K^+ , Mg^{2+} , Ca^{2+} , Cl^- and $(\text{SO}_4)^{2-}$, and salts of these such as halite (NaCl) and gypsum ($\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$) crystallize out. High evaporation rates are common in arid climates; thus, ancient evaporite deposits are a good indicator of ancient arid climates.

evaporites

Summary of Section 4

In this Section we introduced you to only a few of the processes involved in sedimentary rock formation. Once formed, sedimentary rocks may be exposed again at the Earth's surface and undergo a new cycle of erosion and deposition during which their constituent particles become progressively better rounded and better sorted, and the breakdown of their constituent minerals proceeds further.

Many sediments, in fact, contain detrital grains that have experienced several separate episodes of erosion, transportation, re-deposition and erosion. They offer convincing evidence of the cyclical nature of geological processes.

Objectives of Section 4

Now that you have completed Section 4, you should be able to:

- (a) Indicate that you have some understanding of the following terms: physical and chemical weathering, residual minerals, transportation, erosion, denudation, rolling, saltation, bedload, suspended load, Stokes' law, energy of an environment, deposition, sorting, limestone, ooze, evaporite. (SAQs 11–15)
- (b) Relate chemical weathering processes to the mineral compositions of igneous rocks. (SAQs 12, 13 and 15)
- (c) Relate the physical processes of sedimentation in water to natural water environments and to sedimentary rocks. (SAQs 14 and 15; ITQs 5 and 7)
- (d) Appreciate the factors that affect the sorting and rounding of particles in water. (SAQs 14 and 16; ITQ 6)
- (e) Explain the role of the soluble products of chemical weathering in the formation of sedimentary rocks. (SAQ 16)

SAQ 11 Which of the Home Kit specimens S1 (granite) and S4 (gabbro) would you expect to decompose more rapidly as the result of chemical weathering?

- SAQ 12**
- (a) Can you suggest why weathered basalt outcrops usually have a thin, surface layer of brown material?
 - (b) What would you expect to find as the *residual* products from a weathered granite?
 - (c) What would you expect to find as the *residual* products from a weathered gabbro?

- SAQ 13** A river, which is flowing through a narrow gorge cut in rock, has a velocity of 0.3 m s^{-1} and carries a range of particles from clay size upwards. It emerges abruptly from the gorge into a region of flat, low-lying land, and its velocity is reduced suddenly to 0.02 m s^{-1} .
- (a) What range of particle sizes would you expect to find deposited in the river bed where the river emerges from the rock cutting?
 - (b) Would you expect the sediment to be well sorted, moderately sorted, or poorly sorted?

- SAQ 14** Examine Home Kit specimen, S6 (sandstone) closely with your hand-lens.
- (a) Can you suggest in relative terms what the energy of the transporting medium might have been?
 - (b) Can you suggest in relative terms for how long the material underwent transportation?

SAQ 15 In Table 12 are shown the concentrations (as percentages of the total dissolved solids) of various ions and SiO_2 in sea-water and river water. (The river water figures have been corrected for the salts recycled by sea spray). Can you explain why there are such large differences between the two columns?

TABLE 12

| Ion | Concentration in river water | Concentration in sea-water |
|------------------|---------------------------------|-------------------------------|
| Ca^{2+} | 12.5 | 1.18 |
| Na^+ | 1.7 | 30.62 |
| HCO_3^- | 48.9 | 0.40 |
| SiO_2 | 11.0 | negligible |

5 Tectonic and metamorphic processes

In the previous Sections, we have examined two of the essential stages in the rock cycle: the formation of igneous and sedimentary rocks. In this concluding Section, we follow the rock cycle a stage further, and go on to look at the processes that affect rocks once they have been formed.

In particular, we are going to see what happens when rocks are subjected to mechanical forces—what happens when they are squeezed and shoved—and also what happens when they are subjected to extreme physical conditions of temperature and pressure. The two processes together give rise to a whole group of rocks, called the *metamorphic rocks*, that include all those rocks—of either sedimentary or igneous origin—that have been substantially changed since their formation. The changes may be barely perceptible, or they may be extreme, so that wholly new rocks are produced. Both tectonic and metamorphic processes are extremely important to geologists, because they control the final appearance of a rock as it is seen in the field—it is quite unusual to see a rock that shows no sign of *either* process.

5.1 Tectonic processes

Tectonic processes are those concerned with the mechanical deformation of rocks. Almost all rocks are deformed in some way, although this may not always be obvious. As soon as one layer of sediment is deposited on top of another, the lower layer is squashed and compacted, sometimes by as much as 80 per cent of its original thickness. When sedimentary rocks are subjected to sideways compression, by contrast, they may crumple up into much more conspicuous folds. The word *tectonic* is derived from the Greek word for *carpenter*, presumably because a carpenter is responsible for shaping wood into the objects that we finally see and use, such as chairs and tables.

You should recall from Unit 4, Section 3, that when a given stress is applied to a rock, it responds by deforming, and the deformation is measured in terms of *strain*. Four factors control the way in which the rock will be deformed.

Can you suggest what they are?

The most obvious one, of course, is the nature of the rock itself, which is measured in terms of the two elastic moduli that you encountered in Unit 4. Less obvious, but more important for our present purposes, are the *temperature*; whether the stress applied is a *compression* or an *extension*; and *time*. It is common sense to

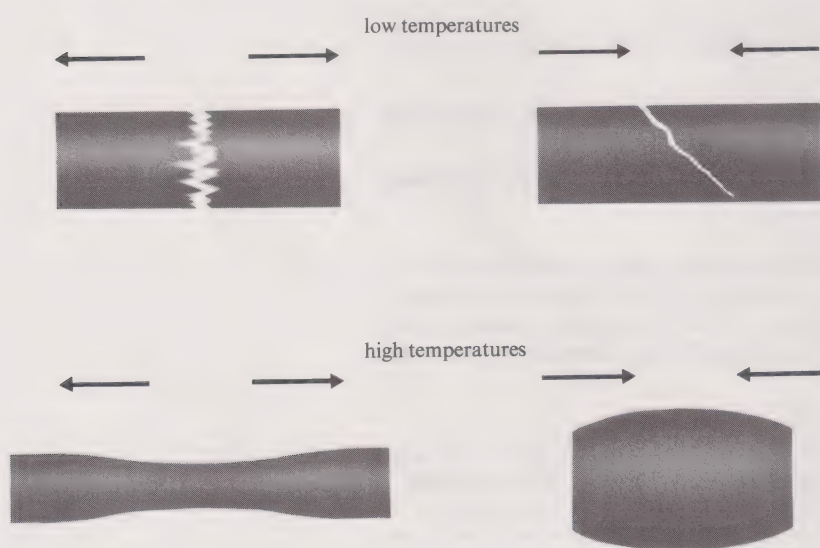


FIGURE 16 At low temperatures, a rock specimen subjected to tension will snap, under compression, it will rupture along an inclined failure surface. At high temperatures, the same specimens are more likely to stretch under tension, or swell up, under compression.

think that if a rock specimen is subjected to an extension, it will eventually break cleanly apart. Under compression, the rock will be ruptured, though not with a clean, smooth fracture (Figure 16). When rocks respond to deformation by snapping or rupturing, this behaviour, not surprisingly, is known as *brittle deformation*.

brittle deformation

At low temperatures, rocks usually behave in this way. At higher temperatures, however, a rock specimen will behave rather differently. Under tension, it will tend to stretch and become thinner rather than snapping apart, whereas under compression it will not rupture, but will merely bulge outwards like a beer barrel (Figure 16). This is called *plastic deformation*, and usually requires temperatures of over 100 °C to bring it about. The higher the temperature, the more easily the rock is deformed.

plastic deformation

Regardless of whether it is under compression or tension, or at high temperatures or low, the effect of *time* is the same. When a rock is stressed, it will respond with an *immediate* deformation. If the stress is maintained continuously, and if it is sufficiently great, the rock will continue to deform indefinitely. This is the phenomenon known as *creep*. Like plastic deformation, creep is accelerated at high temperatures. Creep is important to engineers designing large mechanical structures intended for a long life, and it causes them considerable headaches. But, with a very much longer time-scale, creep is vastly more important in geology, and almost all of the rock structures that you are likely to see are the result of creep deformation.

creep

5.2 Structures produced under tension

Structures produced under tension are perhaps the simplest and most widespread rock structures. When an igneous rock cools, it tends to shrink slightly, and this sets up tensional stresses throughout the whole mass of rock. The rocks respond by breaking up into a well-defined pattern of polygonal fractures known as *joints*, which give the rock a highly distinctive appearance. The Giant's Causeway in Antrim is a magnificent example of this phenomenon (Figure 17). Joints are simply cracks in the rock, where it has split open *without* any movement taking place *along* the crack. They are extremely common in all kinds of rock, and can be seen in almost any outcrop of rock anywhere.

joints

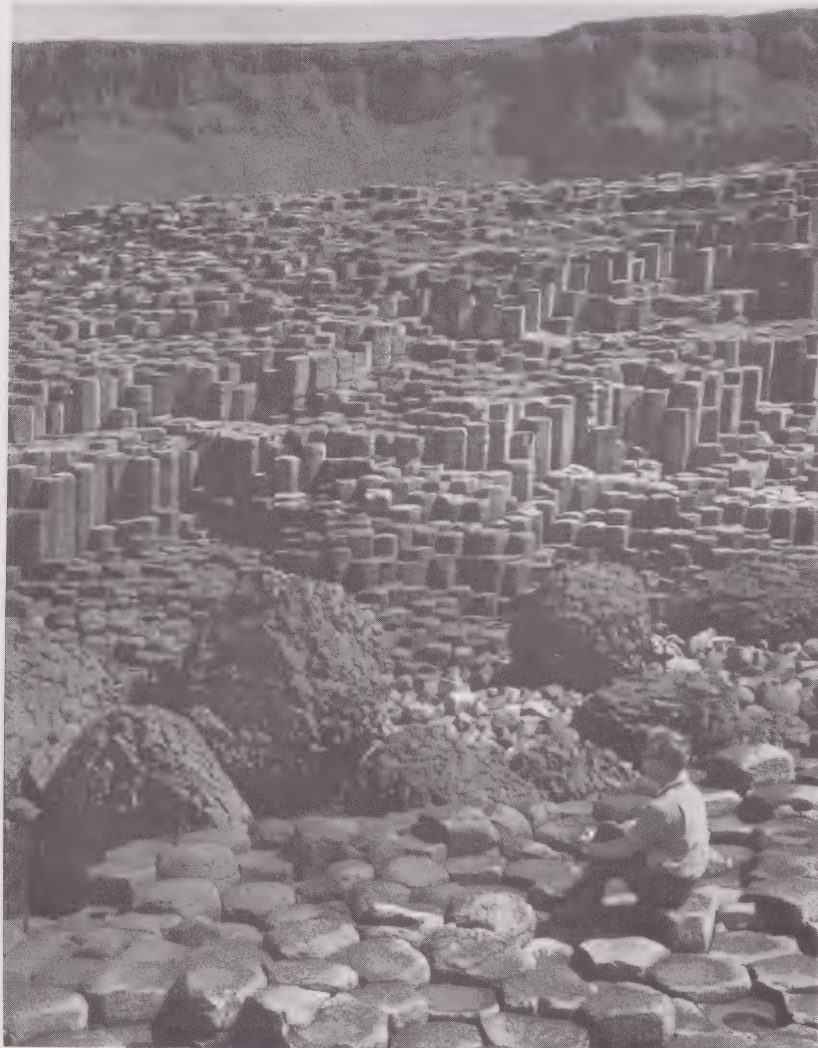


FIGURE 17 The Giant's Causeway, Antrim, showing the myriads of elegant columns produced by cooling within a basalt body. (Institute of Geological Sciences photo)

Qualitatively different from joints are *faults*. These are also planar cracks running through rocks, and are also produced by tensional stresses, but faults are formed when movement takes place along the fracture plane, causing a relative offset between features on different sides of the fault. You were introduced to an important kind of fault in Units 6 and 7, Section 4.9. These were transform faults, a rather complex kind that involve lateral movement. By far the most common faults, however, are called *normal faults*, where one side of the fault is simply dropped down relative to the other (Figure 18a and b).

faults

normal faults

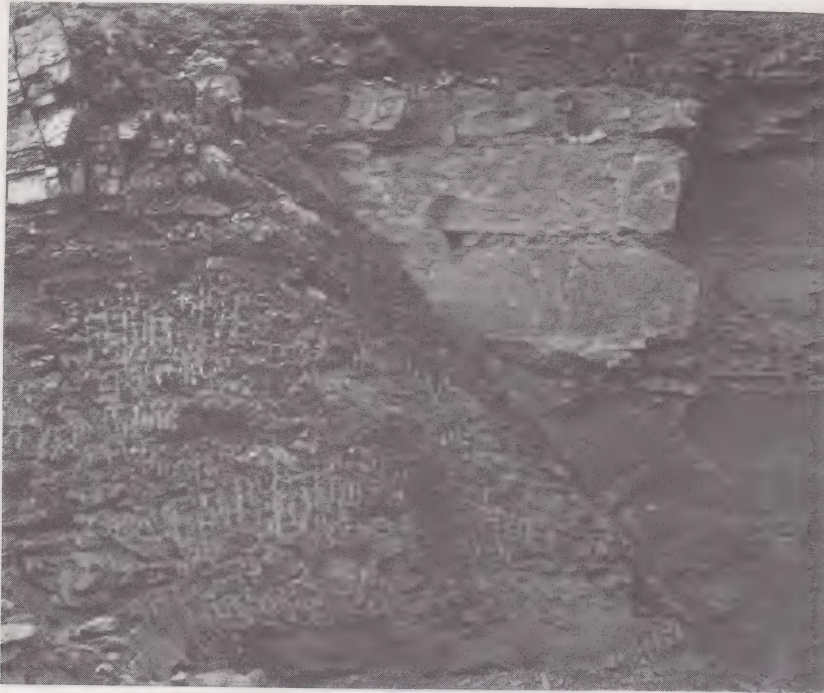
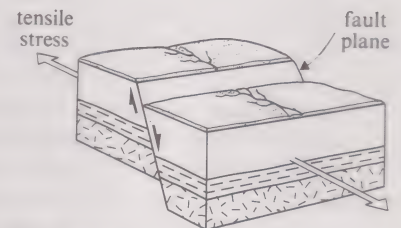
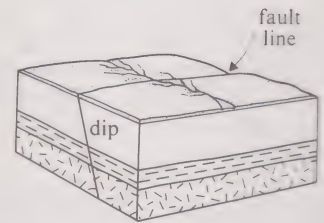


FIGURE 18 (a) A typical small-scale normal fault. The rocks on the right have moved downwards relative to those on the left. The pale-coloured bed can be matched up on either side of the diagonal fault plane. (Institute of Geological Sciences photo)



(b) normal fault

FIGURE 18 (b) Block model illustrating the formation of a normal fault. Compare this with the fault in Figure 18a.

Do you suppose faults would be formed in rocks under tension at high temperature? If not, what do you suppose would happen?

With increasing temperature, rocks would be less likely to break and form faults, and more likely to stretch out plastically, like heated pitch.

5.3 Structures produced under compression

When rocks are squeezed hard, especially at low temperatures, they rupture. The commonest kinds of structure produced in this way are known as *thrusts* (Figure 19), where a large mass of rock breaks in two along a low-angled plane, and the

thrusts

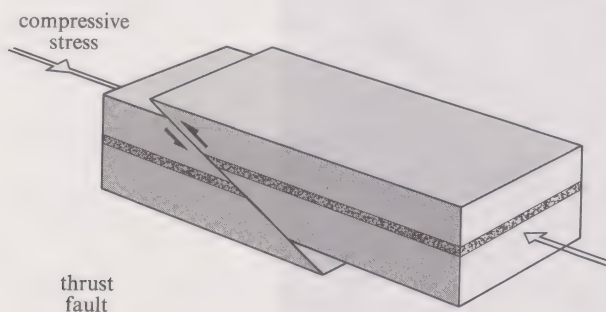


FIGURE 19 Block model of a thrust fault. Notice the difference between this and Figure 18b.

upper block overrides the lower. Some of these moving blocks may be huge, measured in kilometres, and cases are known of thrust blocks moving for tens and even hundreds of kilometres.

When layered sedimentary rocks are subjected to compression, they may form thrusts or, especially when stressed over long periods, they may not break, but pile up into folds instead. The shortening produced absorbs the compressive stress. *Anticlines* and *synclines* are well-known kinds of folds in rocks; they can be seen on a huge scale in mountain chains in the Alps and, on a smaller, but still impressive, scale in many British sea-cliffs and crags (Figure 20).

folds



FIGURE 20 Folds in sedimentary rocks due to compressive stresses. These are at Stair Hole, Lulworth Cove, Dorset. The same stresses that squeezed Dorset to form these folds were also involved in the formation of the Alps. (Institute of Geological Sciences photo)

5.4 Metamorphic processes

Study comment Before reading any further in this Section, you should make sure that you have worked through Band III of Audio-visual tape AC 93, which deals with metamorphic rocks. This will ensure that you are familiar with the rocks and concepts discussed in this Section.

Metamorphic rocks are, as you should be aware by now, rocks that have been derived from other rocks through chemical and physical processes. (The word *metamorphism* has Greek roots, which mean something like 'changing form'.) There is nothing very mysterious about metamorphic processes in rocks, although the details of the processes may be complex. As you heard on the Audio-vision tape, an everyday example of metamorphism is the 'firing' of clay to make pottery; another is the toasting of bread.

The simplest example of metamorphism in geology is *contact metamorphism*, which describes the changes brought about in a rock when it is intruded by a body of hot magma. You will recall that when plutonic rocks are intruded into other rocks, they chill against their surroundings, and a fine-grained chilled margin is produced within them. The surrounding rocks do not escape scot free, and may be quite severely affected by the pluton.

contact metamorphism

What *two* factors are likely to control the results produced?

They are the temperature of the pluton and, less obviously, its size.

The size of the intrusion is important because rocks are such poor conductors of heat that the volume of rock heated up by a small body, such as a 1 m thick dyke, is almost negligible, even though the magma in the dyke may be a thousand degrees hotter than its surroundings. By contrast, a 1 km thick pluton will cool slowly enough to heat up a large volume of rock. It follows that the nearer they are to the pluton, the hotter the surrounding rocks get.

Contact metamorphism has two extremely characteristic effects. First, the surrounding rocks get 'baked'—almost literally. They become extremely hard (like an overcooked cake) and very splintery. Quarrymen of old had a special name for these hard rocks; they called them *hornfels*. (They probably had other names as

well, but these would be unprintable . . .) Second, and more important for our purposes, rocks affected by contact metamorphism often have a spotty, measly appearance. This is due to the growth of *new minerals* within them. The new minerals grow in response to the new physical conditions introduced by the intrusion of the pluton, and the nature of the minerals that grow is directly controlled by the temperature reached in the contact zone around the intrusion (Figure 21).

Contact metamorphism offers an elegant illustration of metamorphic processes—it is easy to see why the rocks were heated, and to trace the variations in the metamorphosed rock around the pluton. Observations of these changes reveal an important fact: the *chemical composition of the metamorphosed rocks is not affected*. The changes that take place are purely *mineralogical*.

A fair analogy can be drawn with a cake baked in an oven—the finished cake looks nothing like the initial mixture that went into the oven, but *nothing has been added to it*. And the only material that has been subtracted is water in the form of steam, and perhaps a little carbon dioxide.

Furthermore, just as the final flavour and texture of the cake depend on the conditions within the oven—especially the temperature—so the minerals that grow in metamorphic processes are controlled by the prevailing conditions of temperature *and* pressure. *Different minerals are characteristic of different sets of physical conditions*.

This key fact enables us to make sense of the most important kind of metamorphism, *regional metamorphism*. As its name implies, this process affects huge areas of rock, often extending over tens of thousands of square kilometres. In fact, most of the rocks making up the continental crust of the Earth are regionally metamorphosed rocks.

Can you suggest where such large masses of rock could be subjected to high temperatures and pressures?

You may recall from Units 6 and 7, Section 5.3, that destructive plate margins, where mountain belts are born, are also characterized by metamorphic processes.

Both constructive and destructive plate margins are characterized by high *heat flow* (Units 6 and 7, Section 4.5). In both of these environments, therefore, one would expect to find deeply buried rocks that have been heated to high temperatures for long periods. We shall concern ourselves here only with what goes on at destructive plate margins.

The rate at which temperature increases with depth in a given region is measured in terms of degrees Celsius per kilometre. This is known as the *thermal gradient*.

Figure 22 shows the variation of temperature with depth for two quite different areas in the United States: one on the edge of the Rockies, the other right in the middle of the continent.

What are the thermal gradients in each case, and what is the temperature at a depth of 20 km in each case?

The thermal gradient beneath the stable, continental part of the United States of America is nearly 15 °C per km, but it is 25–30 °C per km beneath the mountain belt. Thus the temperature at a depth of 20 km is 600 °C beneath the Rockies, twice that beneath the continental part of the United States of America.

It's clear, then, that one does not have to go far down beneath the Rockies to encounter high temperatures. So what happens to the rocks down there? Well, we cannot burrow down that far beneath the Rockies to see for ourselves, but we *can* look at the roots of older mountain belts that have been brought to the surface. In areas such as the Highlands of Scotland, where rocks involved in mountain building events over 600 million years ago are exposed, we can trace exactly what happens to rocks that started off life as ordinary, wet sediments, the kind of thing you might find forming today on a beach or in a river.

Consider mud. When fresh, mud has properties that scarcely need description—it is wet, soft and squishy. When deposited in substantial thickness in, say, an estuarine environment, muds experience a number of changes as water is squeezed out



FIGURE 21 Spotty, contact metamorphosed rock. The spots in this rock were formed by metamorphism near the contact of a granite in North Wales.

regional metamorphism

thermal gradient

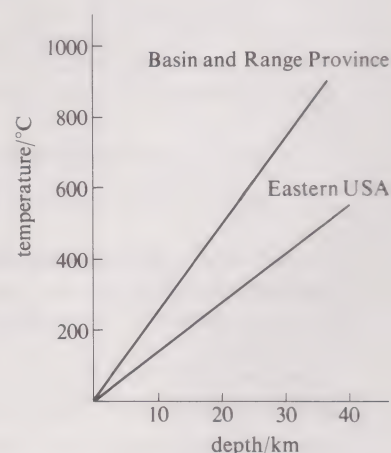


FIGURE 22 Variation of temperature with depth for two areas within the USA. The Basin and Range Province is in the foothills of the Rockies.

of them and they become hardened. Eventually, the mud becomes *lithified*, and a rock results. These changes, however, do not warrant the term *metamorphism*. This is reserved for the changes that take place in response to heat and pressure. If subjected to a mild degree of metamorphism, our 'mud' will grow new minerals, usually flaky, micaceous ones, and will acquire a perceptible lamination. Your specimen S8, the phyllite, was produced by exactly this kind of metamorphism. The phyllite started off life as a muddy sediment, but sufficient new mineral growth has taken place within it to give the rock a distinctive sheen, and an excellent lamination.

The *schist*, specimen S9, is the next stage in the process. It is much more coarsely crystalline, and it is difficult to see much of its original nature. Although it may once have been a sediment similar to that from which the phyllite formed, all trace of sedimentary properties has gone. The *gneiss* (S10) is the last step in the chain; once a rock has reached gneiss grade, it is very close to becoming a granite. As you can see from your specimen, the texture of the gneiss is much more like that of the granite than a phyllite; the only guide to its metamorphic nature is the well-displayed planar banding.

The word 'granite' should have suggested something important to you. Granites are *igneous* rocks. We have been dealing with metamorphosed rocks in this Section, so you should now begin to see a link emerging between the most highly metamorphosed rocks and igneous rocks.

Refer back to Figure 22 and look once more at the thermal gradient beneath the Rockies. What is the temperature at 40 km depth?

It is about 1000 °C.

The importance of this is that, because granites have low melting points—much lower than basalts—one can expect to find granite magmas at depths greater than about 40 km below the Rockies. Hence, in the extreme conditions of regional metamorphism, sediments are brought up to temperatures that are sufficiently high to melt them and for granitic magmas to be produced.

It is important to emphasize that wholesale, complete melting of sediments does *not* take place. Just as *partial* melting of mantle peridotite forms basalt (Section 3.2), so partial melting of the metamorphosed rocks is all that is needed to produce granitic magmas. In general, only a few per cent of the total is melted, and this, of course, is the granitic component. The other, higher melting point components remain behind.

In the deep roots of mountain chains, then, igneous and metamorphic processes converge. In general, however, large volumes of granite are not produced by simple melting of sediments. It is thought that most granite magmas are produced at destructive plate margins, where andesitic magmas are produced by partial melting of mantle material, and undergo successive stages of 'refining' to produce more siliceous melts. A relatively small proportion of material from the lower continental crust, possibly of sedimentary origin, also gets melted and contributes to the granitic magmas.

The magmas that are produced move upwards and away from their source in the lower crust, and soon lose all trace of the contribution that melted sediments have made to them. They accumulate to form large plutons, and as such are wholly igneous rocks.

Thus the rock cycle is completed: igneous rocks are eroded to form sediments, the sediments are metamorphosed and may be eventually melted to contribute towards a fresh generation of igneous rocks. This cycle is extremely important, so it is worth while summarizing all the steps that are required to go from mantle peridotite to granite:

- 1 Partial melting of mantle peridotite at destructive plate margin produces andesitic magmas, which form island arcs.
- 2 Sustained igneous activity produces more siliceous magmas, including granites, from the mantle.
- 3 Andesitic and granitic igneous rocks are exposed at the surface, are eroded, and form fresh sediments.
- 4 Sediments are deeply buried and subjected to regional metamorphism at destructive plate margins.

- 5 Regional metamorphism produces progressive changes in sediments: from sediment, through phyllite and schist to gneiss.
- 6 Eventually, partial melting of gneisses produce new granitic magmas, which are incorporated into andesitic and more siliceous magmas that rise from the subduction zone.
- 7 New granitic magmas move away and upwards in the crust to become igneous intrusions.
- 8 Cover rocks are eroded, and igneous intrusions are eroded to form fresh generation of sediments.

5.5 The rock cycle—a concluding comment

A unifying theme in this Unit has been the rock cycle: the ceaseless erosion of rocks, the transport of sediment by rivers, and the deposition of sediment to form new rocks. Parts of this cycle are so well known that they have become entrenched in popular literature. Travel brochures are full of ‘brooding mountains sculpted by Nature through long aeons of ceaseless toil’.

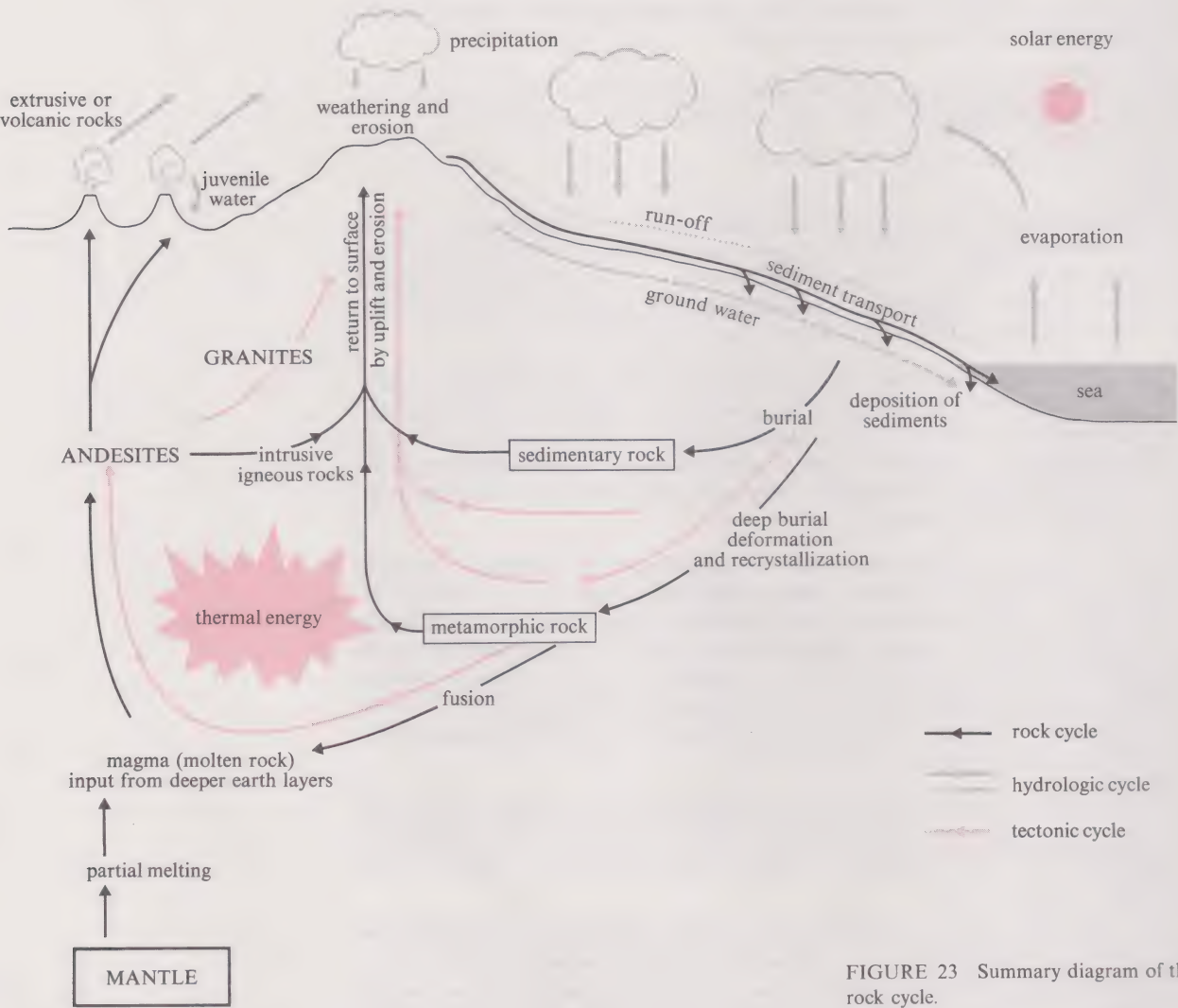


FIGURE 23 Summary diagram of the rock cycle.

The extreme consistency of the geological processes involved in the rock cycle led, in the nineteenth century, to the formulation of a key concept in geology, that of Uniformitarianism (Unit 26, Section 3.4). Proponents of this concept argued that the processes we see operating on Earth today are the same as those that have always operated, and that we can understand old rocks by interpreting them in

terms of processes operating today. This concept holds up extremely well in relation to the rock cycle, with one important exception.

The exception lies in the formation of continental crust from mantle peridotite. The rock cycle, as we know it, simply could not get under way without the formation of the continents. These, as we have seen, are formed by the *irreversible* formation of andesites from mantle peridotite and, because the process *is* irreversible, there is a slow and steady increase in the amount of continental crust. There is, in effect, therefore, a steady input of new material into the rock cycle. Once andesites have been formed, however, they become part of the cycle, and the other processes remain unaffected.

Objectives of Section 5

When you have finished this Section, you should be able to achieve the following Objectives:

- 1 Summarize the different ways in which rocks can be deformed. (SAQs 16 and 17)
- 2 Describe, and explain the reasons for the formation of joints, faults and folds. (SAQs 16 and 17)
- 3 Describe the nature of metamorphism, and explain the relationship between metamorphism and thermal gradient. (SAQs 18 and 19)
- 4 Summarize the progression of different rock types produced in regional metamorphism. (SAQs 18 and 19)

SAQ 16 Which of the following statements correctly describe tectonic processes?

- (a) Rocks are stronger at high temperatures than at low.
- (b) Rocks may break or fracture when subjected to both extensional and compressive stresses.
- (c) Creep will not take place in rocks unless a certain minimum stress is applied.
- (d) The rate at which creep takes place is independent of other factors.
- (e) Rocks are stronger in tension than they are in compression.
- (f) Joints in rocks are always the result of compressive stresses.

SAQ 17 What kinds of structures might be produced in rocks by compressive stresses at low temperatures; and what structures would be more likely at high temperatures? Give reasons for your answers.

SAQ 18 If granitic rocks melt at a temperature of 850 °C, at what depth would you expect to find partial melting of crustal rocks taking place in a region where the thermal gradient is 15 °C per km? You may assume that the surface temperature is 20 °C.

SAQ 19 Match each of the following rock types (1–4) with one of the sets of conditions in the key (A–D).

Rocks

- 1 Partially melted sediment.
- 2 Phyllite.
- 3 Hornfels.
- 4 Gneiss.

KEY

- A Contact metamorphism.
- B High-grade regional metamorphism.
- C Thermal gradient of 28 °C per km extrapolated to 30 km.
- D Low-grade regional metamorphism.

Objectives

- 1 Show that you have some understanding of the following terms: mineral, silicate mineral, crystallization temperature, atomic structure, physical and chemical weathering, residual minerals, transportation, erosion, denudation, rolling, saltation, bedload, suspended load, Stokes' law, energy of an environment, deposition, sorting, limestone, ooze, evaporite.
- 2 Relate the colour of a silicate mineral to its chemical composition, and the chemical composition to structure and to temperature of formation of the mineral.
- 3 Relate the chemical composition and structure of silicate minerals to their densities.
- 4 Make predictions about mineral composition, temperature of formation, rate of cooling, order of mineral crystallization and density of an igneous rock from a hand specimen.
- 5 Describe the possible sources of the Earth's internal heat.
- 6 Explain systematically, and solve simple problems based upon the ways in which the compositions of igneous rocks can be changed, and new rocks formed, by partial melting and fractional crystallization.
- 7 Summarize the main differences between volcanic and plutonic rocks.
- 8 Predict the heights to which volcanoes will grow in given circumstances.
- 9 Explain how continents are formed by destructive plate margin processes.
- 10 Describe different ways in which rocks of granitic composition may be formed.
- 11 Relate chemical weathering processes to the mineral compositions of igneous rocks.
- 12 Relate the physical processes of sedimentation in water to natural water environments and to sedimentary rocks.
- 13 Appreciate the factors that affect the sorting and rounding of particles in water.
- 14 Explain the role of the soluble products of chemical weathering with formation of sedimentary rocks.
- 15 Summarize the different ways in which rocks can be deformed.
- 16 Describe, and explain the reasons for the formation of joints, faults and folds.
- 17 Describe the nature of metamorphism, and explain the relationship between metamorphism and thermal gradient.
- 18 Summarize the progression of different rock types produced in regional metamorphism.

ITQ answers and comments

ITQ 1 (a) Ruby is a mineral. It occurs naturally, has a crystalline structure, and a fixed chemical composition. It is also, of course, a gemstone.

(b) Cooking salt is not strictly a mineral. This is a bit of a trick question, because salt *does* occur as a natural mineral, but cooking salt is almost invariably an industrially recrystallized form.

(c) Industrial diamonds may be found naturally, or they may be manufactured—so some are minerals and some are not!

(d) Snowflakes are strictly speaking minerals! They have all the necessary properties of crystalline form and chemical composition. They are never, however, listed in textbooks of mineralogy.

ITQ 2 (a) High-temperature forms are magnesium-rich, whereas low-temperature forms are iron-rich.

(b) High temperature plagioclase feldspars are calcium-rich, whereas low-temperature plagioclase feldspars are sodium-rich.

ITQ 3 The mass of a 1-metre-square column of lithosphere would be 1.92×10^8 kg. (thickness \times area \times density). A basalt column of the same mass would be 66.6 km high (1.92×10^8 divided by the density of basalt). Thus the volcano could reach a height of 6.6 km above the ocean floor.

ITQ 4 There is a progressive *increase* in SiO_2 , Na_2O and K_2O in going from peridotite to granite, and a *decrease* in MgO and $\text{Fe}_2\text{O}_3 + \text{FeO}$. The implication is that, starting with a liquid of *peridotite* composition, rocks of any intermediate composition up to granite could be made by subtracting the right elements, particularly MgO and FeO . The minerals required to make this subtraction need not concern us here.

ITQ 5 From Figure 14 you can see that water-current velocities must drop to 0.01 m s^{-1} , or 1 cm s^{-1} before fine sand will be deposited.

ITQ 6 Wave action in sheltered bays is fairly constant. Coarse material will have been deposited already around the headlands and fine material will be constantly winnowed out by the wave action, so you would expect beach sands in sheltered bays to be well sorted.

ITQ 7 The current velocity required to erode a 0.125 mm sand grain is approximately 20 times that required to deposit it.

SAQ answers and comments

- SAQ 1** (a) Your completed Table 7 should look something like Table 12. Note that in the granite (S1) you would expect feldspars and micas to crystallize out around the same time; and—in the gabbro—pyroxenes and feldspars will crystallize out at the same time.
- (b) The colour difference is a result of the mineral proportions. The granite (S1) contains only 10 per cent of dark ferromagnesian minerals and therefore appears pale coloured. The gabbro (S5) and peridotite (S4) contain 40 per cent and 100 per cent ferromagnesian minerals respectively, and so appear much darker coloured.
- (c) The iron and magnesium rich minerals olivines and pyroxenes have the simplest structures with the least oxygen sharing between the $(\text{SiO}_4)^{4-}$ units. Therefore they have close packed structures and high densities. A rock like S4 which contains all olivine and pyroxene will therefore have a high density. A rock like S1 which contains 90% feldspars and quartz with open structures will have a low density. S5, the gabbro, contains almost 50% of each type of mineral and could therefore be expected to have a density somewhere in between S1 and S4. Note that we cannot comment on the effects of the relative Fe and Mg contents of the olivines and pyroxenes in detail from these hand specimens.
- (d) The feldspar crystals were able to grow unimpeded from the magma and so formed good crystal shapes. The quartz was the last mineral to crystallize and had to fill in the spaces left between the other minerals and so is irregular in shape.

- (e) As all three samples are coarse grained, this implies that they all cooled and crystallized slowly, probably at depth below the Earth's surface.
- SAQ 2** (a) The overall colour of the rock is very dark, which suggests it must have a high proportion of iron- and magnesium-rich minerals.
- (b) As iron- and magnesium-rich minerals all form at relatively high temperatures, it is likely that this rock formed from a magma at a high temperature.
- (c) As the rock is fine grained, it cooled rapidly and so crystallization occurred rapidly, with many crystals growing at once in the magma as it chilled below their crystallization temperatures.
- (d) The density of S3 was between that of S1 and S4, and should have a density similar to that of S5, the gabbro.

It should be clear from this exercise that gabbro (S5) and basalt (S3) have quite a lot in common: their densities are similar, and their dark colours suggest that both contain a substantial proportion of iron- and magnesium-rich minerals. Thus, although we cannot see the individual minerals in the basalt, we can deduce that it has a broadly similar mineralogy to the gabbro. Since the gabbro is coarse grained, and the basalt fine grained, they could have been derived from the same type of magma, if the gabbro cooled and crystallized slowly, and the basalt cooled and crystallized rapidly.

TABLE 12 Suggested answer to SAQ 1

| Rock sample | Average mineral composition | Colour of mineral | Overall rock colour (light or dark) | Relative density (high, moderate, low) | Relative temperature of formation (high, moderate, low) | Anticipated sequence of mineral crystallization |
|---------------|--------------------------------|--|-------------------------------------|--|---|---|
| S1 GRANITE | 10% biotite mica | brown-black | light coloured | low | low | mica + feldspars |
| | 60%, mainly Na- and K-feldspar | pinkish-white | | | | ↓ |
| | 30% quartz | colourless | | | | quartz |
| S5 GABBRO | 10% olivine | difficult to see any—may be occasional green brown cystals | dark coloured | moderate | moderate | olivine |
| | 30% pyroxene | dark greeny-black | | | | ↓ |
| | 60% Ca- and Ca-Na-feldspar | white | | | | pyroxene + feldspar |
| S4 PERIDOTITE | 85% olivine | olive green | dark coloured | high | high | olivine |
| | 15% pyroxene | dark greeny-black | | | | ↓ pyroxene |

SAQ 3

A Not true. Although it is correct that the crust contains a relatively high proportion of radioactive elements, these cannot account for high temperatures in the mantle and core.

B Not true. Clearly, the amounts of radioactive elements present today are only a fraction of what they were initially.

C Not true. Lead by itself is not radioactive, although it may be formed by radioactive decay of uranium.

D True to some extent. But the Sun's heat has a minimal effect on the Earth's heat balance.

E Probably true. Rates of conduction are so slow that a solid Earth would cool very slowly. Convection currents, however, offer a means of transferring heat rapidly from the centre outwards, so these may have been responsible for relatively rapid cooling.

F Uncertain. Because heat escapes so slowly, it is possible that enough heat is generated within the mantle to keep it mobile.

G Theoretically true. But this heat is of a secondary kind—heat is required in the first place to start the cells moving.

SAQ 4 The components removed are clearly SiO_2 , Al_2O_3 and CaO . But since the andesite magma has very nearly the same SiO_2 content, there will be only a trivial depletion in the residual liquid. Al_2O_3 , by contrast, will be much more heavily depleted, and CaO rather less so. If the plagioclase contained less CaO than the 8 per cent present in andesite magma, then, of course, the residual liquid would be relatively enriched in CaO .

SAQ 5 Only G and I apply *only* to volcanic rocks, all the others could apply to plutonic rocks. Recall the sets of criteria listed in Sections 3.2 and 3.4.

SAQ 6 The weight of a 1-m^2 column of the lithospheric plate is given by the area times thickness times density:

$$\begin{aligned}\text{weight} &= 1 \times 70 \times 10^3 \times 3.3 \times 10^3 \\ &= 70 \times 3.3 \times 10^6 \\ &= 2.31 \times 10^8 \text{ kg}\end{aligned}$$

The thickness t of a column of granite magma 1 m^2 , having the same weight is given by:

$$1 \times t \times 2.6 \times 10^3 \text{ kg}$$

The two weights must be equal, therefore:

$$\begin{aligned}t &= \frac{2.31 \times 10^8}{2.6 \times 10^3} \\ &= 88.8 \text{ km}\end{aligned}$$

The height of the volcano is, therefore, $18.8 - 70 = 18.8 \text{ km}$. There are no volcanoes as high as this on Earth, but there are some on Mars!

SAQ 7 Each kilometre of destructive plate margin erupts $2.5 \times 10^{-5} \text{ km}^3$ of magma each year. The total world production is therefore 50 000 times greater than this, namely 1.25 km^3 . But only 80 per cent of this is *new*, derived from the mantle. Therefore, the volume of new continental crust generated each year is $1.25 \times 80/100 = 1 \text{ km}^3$ per year.

SAQ 8 It must originate from partial melting of pre-existing rocks overlying the subduction zone. These may be andesites, or granites, or sediments that have been formed at an older plate margin.

SAQ 9 Steady removal of olivine and pyroxene crystals from a magma of initially basaltic composition will yield a residual liquid of increasingly granitic composition.

SAQ 10 Granites may be formed in small volumes at *constructive* plate margins by extreme *fractional* crystallization of *basaltic* magmas, but they are formed in much larger volumes at *destructive* plate margins, where progressive episodes of magmatism lead to the evolution of progressively more *siliceous* liquids. Heat carried up into the crust may also cause partial *melting* in the upper crust, thus contributing further to the volume of granitic magma.

SAQ 11 The granite is composed mainly of pink orthoclase feldspars, quartz and biotite mica, which are low-temperature, stable silicates and, therefore, fairly resistant to chemical weathering. The gabbro is composed mainly of calcium-rich plagioclase feldspars and pyroxene, which are high-temperature, unstable silicates and, therefore, not very resistant. You would expect from this that the gabbro should decompose more rapidly than the granite.

SAQ 12 (a) Basalts contain 30 per cent or more iron- and magnesium-rich minerals and, when these weather, iron oxides are left as residual products. These form the brown surface colouration of weathered basalts.

(b) The granite consists of feldspars, micas and quartz. The feldspar will weather to leave residual *clay minerals*; the micas will also weather to *clay minerals* if chemical weathering is prolonged: quartz is not attacked and remains as a residual mineral. So the residual products are most likely to be clay minerals and quartz, and possibly some undecomposed mica.

(c) The gabbro consists of plagioclase feldspars and pyroxene with perhaps a little olivine. The feldspars will weather to clay minerals and the ferromagnesian minerals will completely decompose, leaving only an iron oxide residue. So the residual products are most likely to be clay minerals and iron oxide.

SAQ 13 (a) Using Figure 14, you would expect to find particles within the range 5 mm (maximum grain diameter carried at a current velocity of 0.3 m s^{-1}) to about 0.25 mm (maximum grain diameter carried at a current velocity of 0.02 m s^{-1}):

(b) As the drop in current velocity will have been rapid, we should expect all material within the size range 0.25 mm to 5 mm grain diameter to be deposited. However, as the range of grain sizes is not very extensive, we should expect the sediment to be fairly well sorted.

SAQ 14 (a) You should look for the largest grain size present because this must represent *at least* the largest grain size the transporting medium could carry. In some specimens there may be small pebbles between 0.5 and 1 cm diameter indicating a moderately high-energy environment.

(b) Except for larger fragments such as small pebbles, which become rounded quite rapidly, most of the quartz grains still show angular and irregular surfaces—they are far from well rounded. This suggests that they cannot have been in transportation for long periods. You may also remember from the Audio-vision sequence for Unit 4 that occasional feldspar grains are present. As chemical weathering must continue during transportation, this also suggests that transportation time could not have been extensive, otherwise the feldspar grains would have completely decomposed.

SAQ 15 Ca^{2+} and HCO_3^- are removed from sea-water by marine organisms to form calcareous shells. Similarly, diatoms and radiolaria remove SiO_2 from sea-water. This means that the *relative* proportion of Na^+ in sea-water must be much greater than in river water and, consequently, the concentration of Na^+ in sea-water as a percentage of the total dissolved solids must also increase.

SAQ 16 (a) Incorrect. Rocks are more easily deformed at high temperatures, hence they are *weaker*.

(b) Correct. Although the fracture patterns are different in each case, rocks do break under both compressive and tensional stresses, especially at low temperatures.

- (c) Correct. If this were not the case, structures built of stone, such as bridges and cathedrals would not be capable of supporting themselves for centuries, and would soon creep to destruction!
- (d) Incorrect. An increase in temperature causes a marked increase in the rate of creep.
- (e) Incorrect. Rocks are much stronger in compression. Again, if this were not the case, it would be difficult to make buildings stay up, as almost all the stresses in a building are compressive ones due to the weight of the structure.
- (f) Incorrect. Joints are almost always the result of tensional stresses.

SAQ 17 At low temperatures, rocks are likely to be brittle, hence *thrusts* would be produced. At higher temperatures, more plastic behaviour is likely, and hence *folds* would be produced.

SAQ 18 Since the surface temperature is 20°C , we need to increase the temperature by 830°C . At 15°C per km, this requires $830/15\text{ km} = 55\text{ km}$.

SAQ 19 1C; 2D; 3A; 4B.

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